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Editor-in-Chief
Boris S. Sokolov
Member of the Russian Academy of Sciences,
Counsellor of the Presidium of the Russian Academy of Sciences

Address for correspondence: Staromonetnyi per. 22, Moscow, 109180 Russia
Phones: 7 (095) 951-2164; Fax: 7 (095) 233-5590.

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The Lower Cambrian Boundary in Inner Areas of the North Siberian Platform

V. V. Khomentovskii*, A. B. Fedorov, and G. A. Karlova***

** Joint Institute of Geology, Geophysics, and Mineralogy, Siberian Division, Russian Academy of Sciences, Universitetskii pr. 3, Novosibirsk, 630090 Russia*

*** Siberian Research Institute of Geology, Geophysics, and Mineral Resources, Krasnyi pr. 67, Novosibirsk, 630104 Russia*

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Abstract—The Vendian-Cambrian sequences recovered by boreholes K-203 and K-204 and by the deep Eiik Hole 343-0 in the North Siberian platform yielded abundant small-shelled fossils characteristic of the *N. sunnaginicus* Zone of the Tommotian Stage and also of *P. antiqua* and *A. trisulcatus* zones of the Nemakit-Daldynian Stage. These sequences appear to be more fossiliferous than the coeval sequence of the reference Markha borehole, which is used for the correlation of natural exposures in the northern part of Siberia with the borehole sections in its inner part. New data enable us to use the recognized stratigraphic subdivisions in the inner oil-and-gas provinces of the Siberian platform and to outline the elaboration method of the general stratigraphic scale for the Vendian-Cambrian boundary beds.

Key words: *Vendian, Cambrian, Nemakit-Daldynian Stage, Tommotian Stage, small-shelled fauna, the Precambrian-Cambrian boundary, Siberia.*

Difficulties in extending the formally approved Newfoundland model for the lower Cambrian boundary (for discussion and bibliography see Rozanov *et al.*, 1997) require us to use the traditional Siberian model of this boundary as corresponding to the base of the *Nochoroicyathus sunnaginicus* Zone or to that of the Tommotian Stage (*Resheniya...*, 1989). The position of this boundary in the inner areas of the Siberian platform is still ambiguous.

The lowest level of the boundary was assumed to be marked by the substitution of the so-called second assemblage of organic-walled microfossils by the third one (Volkova *et al.*, 1980). This level was considered as occurring within the Kursovskii Formation (see below), the upper part of which was assigned to the Ayanian Horizon of the Lower Cambrian (*Resheniya...*, 1983; Faizulina *et al.*, 1982; Rudavskaya and Vasil'eva, 1985). However, it soon became clear that the composition of acritarch assemblages was strongly controlled by sedimentation environments, and the Ayanian boundaries turned out to be diachronous (Khomentovskii *et al.*, 1987). At present, the substitution of *Leiospheridia* assemblage II by assemblage III with acanthomorphic forms has proved to be considerably dependent on the basin depth, and recurrent assemblages can be recorded in the section (e.g., Zang and Walter, 1992). The Vendian age of the acritarchs from the Ayanian Horizon is beyond doubt now (Faizulina, 1996).

The positioning of the lower Cambrian boundary on the basis of small-shelled fossils (SSF) in interiors of the Siberian platform also is difficult and ambiguous, because, in the transitional facies region of the Siberian platform, the base of the Tommotian Stage is marked by an abrupt diversification of the SSF assemblage and by the appearance of the archaeocyathid macrofauna (Fig. 1), whereas in the eastern facies region the Vendian assemblage shows very gradual enrichment in typical Tommotian forms. In the last case, some researchers determine the lower Cambrian boundary by the appearance of several characteristic Tommotian taxa, but others believe that its position should be substantiated by the occurrence of all the Tommotian forms known in the transitional facies zone. Adherents of the second standpoint recognize several pre-Tommotian subdivisions of the Lower Cambrian such as the Khayalakhian (Val'kov, 1987) and Manykaian (Missarzhevskii, 1989) stages. Comprehensive biostratigraphic investigations of well-exposed sections in the southern and northern Siberian platform (Fedorov and Shishkin, 1984; Khomentovskii and Karlova, 1992) showed that SSF assemblages compatible in composition with the earliest assemblage of the *N. sunnaginicus* Zone of the transitional area were recorded only in the *Dokidocyathus regularis* Zone or even in the Atdabanian Stage of the eastern facial region. At the same time, the facies-independent appearance level of characteristic Sunnaginian forms is synchronous over an enormous territory (Khomentovskii and Karlova, 1992, 1994).

Positioning of the lower Tommotian boundary in the stratigraphic scale of the inner areas of the Siberian platform is a complex problem, since this scale is com-

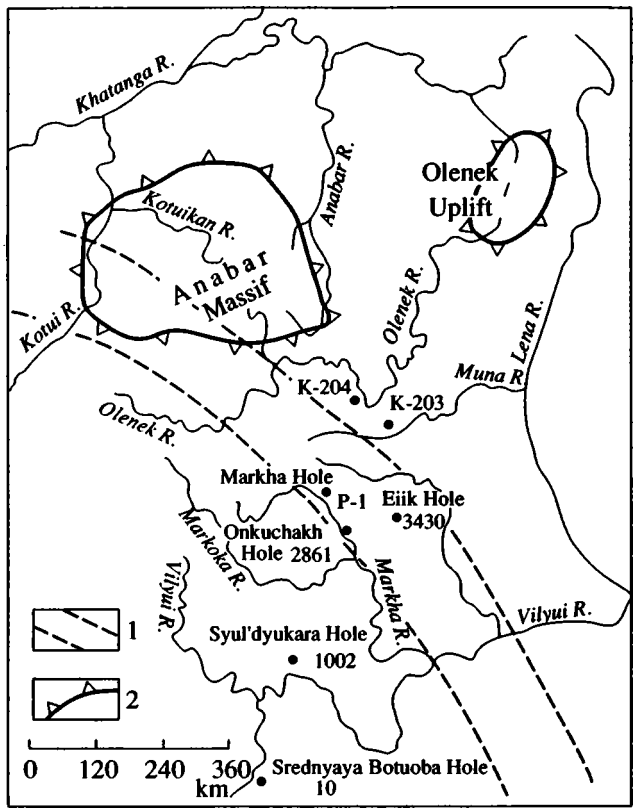


Fig. 1. General map of the territory under study: (1) position of the transitional facies zone in the Vendian-Cambrian time; (2) area of the exposed Vendian-Cambrian deposits.

piled on the basis of logging data for exploration boreholes and has inadequate paleontological characteristics. The actual core recovery from most exploration boreholes does not exceed 5–7% and characterizes only limited stratigraphic intervals. As a result, the recovered fossils are scarce despite the great number of boreholes, and it is possible to detect their successive assemblages in particular sections only extremely rarely.

In this respect, a section of the Markha borehole (Gritsik, 1969), which yielded a series of logging and paleontological data (see Fig. 1), was always considered as very important. However, noticeable facial changes traceable from this hole to the nearest exposures of Vendian-Cambrian deposits of the Anabar massif located 300 km away resulted in several variants of correlation. Some researchers considered as a very significant marker horizon the limestone member containing calcareous algae (depth interval 1630–1640 m) and correlated it with the Koril Member of the Anabar section. They placed the lower Tommotian boundary immediately above this unit, citing the occurrence of *Oelandiella sibirica* Miss. in the interval of 1608–1612 m (Kolosov, 1977; Shishkin and Zhuravleva, 1990; and others). They also correlated the base of the Manykai Formation with a member of sandy dolomites (interval 1815–1835 m). Other researchers (Levi, 1978; Makarov, 1966; Khomentovskii,1985; and others) emphasized the significance of a member of gravelly sandstone (depth 1700 m) and overlying greenish-gray marl (Fig. 2) for the definition of this boundary. In this case, as in the western Anabar region, *Oelandiella* species was thought to occur in the pre-Koril member and assigned to the *P. antiqua* Zone (Khomentovskii and Karlova, 1992). The Koril Member was subsequently correlated to the algal limestone (interval 1570–1600 m). It should also be noted that identifications of SSF from the Markha Hole, which were made more than 30 years ago, need to be revised in order to establish reliably the level of the lower Tommotian boundary in the region.

Boreholes K-203 and K-204 in the interfluve between the Muna and Olenek rivers provided new data elucidating the correlation between the Vendian-Cambrian sequences of the Anabar massif and inner areas of the Siberian platform, because, in contrast to the Markha borehole, they are located near the bedrock exposures in the same (eastern) facies zone (Fig. 1).

Rock sequences in these boreholes include units evidently analogous to the Staraya Rechka and Manykai formations of the Anabar Uplift. The former formation unconformably resting on metamorphic rocks of the basement is composed of gray dolomites, some beds and members of which are enriched with clastic material. This feature permits the detailed correlation of these sequences with each other and (according to geophysical logging data) with deep borehole sections in the south (Fig. 2). In borehole K-203, the Staraya Rechka Formation shows a high content of anhydrite in the lower part and abundant stromatolitic dolomites in the upper portion. The boundary between these parts is coincident with the base of a sandy dolomite member in the depth interval of 1050–1075 m.

The Manykai Formation (interval 866–1000 m) is composed of gray and greenish-gray clayey limestones and marls. Its lower part, which is 40 m thick, mostly consists of clastic carbonate rocks, which vary in composition from dolomite to limestone and in clast size from fine-grained sandstone to gritstone and even to small-pebble conglomerate.

The main upper part of the Manykai Formation is composed of clayey-carbonate rocks. The formation contains small-shelled fossils throughout its interval.

There is a difficulty related to the precise positioning of the lower boundary of the Emyaksy Formation in the sequence under discussion. Its thick lower part is also composed of gray clayey limestones and dolomites, whereas the proportion of red beds characteristic of the formation gradually increases upsection. We arbitrarily place this boundary (depth 863 m) at the level separating assemblages of the *Purella antiqua* and *Nochoroicyathus sunnaginicus* zones, the latter typical of the basal Tommotian Stage (see below).

The Staraya Rechka Formation recovered in borehole K-204 shows the same composition, structure, and thickness as in borehole K-203. In contrast, the Many-

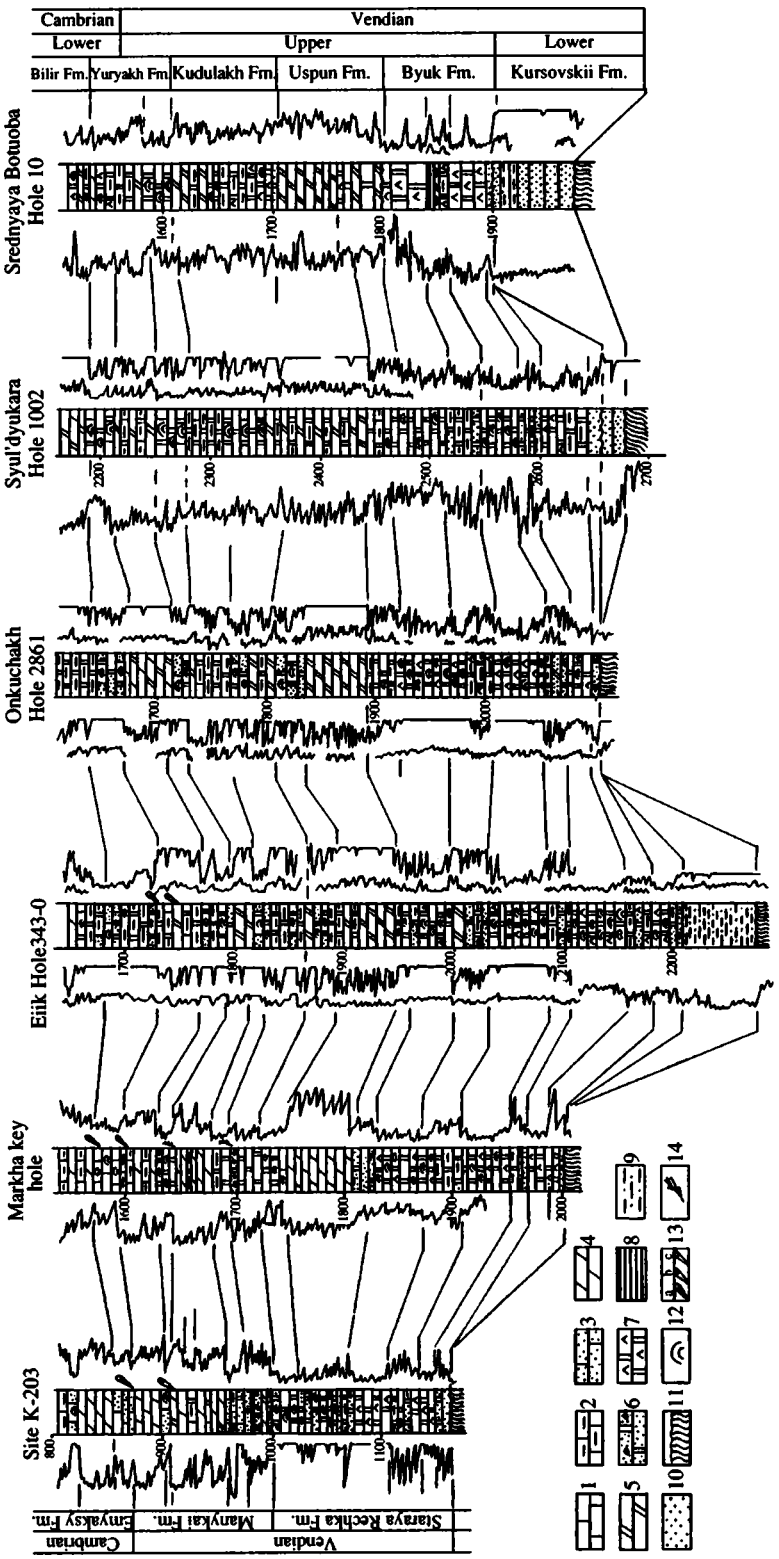


Fig. 2. Correlation of Vendian deposits based on gamma-logging in boreholes in the northern Siberian platform: (1) limestone; (2) clayey limestone; (3) sandy limestone; (4) marl; (5) dolomitic marl; (6) sandy dolomite; (7) anhydrite dolomite; (8) mudstone; (9) siltstone; (10) sandstone; (11) crystalline basement; (12) stromatolites; (13) small-shelled fossils of (a) *N. sunnaginicus*, (b) *P. antiqua*, and (c) *A. trisulcatus* zones; (14) calcareous algae.

kai Formation significantly changes: its lower member is substituted by clayey deposits. To the south and southwest, the analogues of the Manykai Formations demonstrate a gradually decreasing content of clastic material. The Markha reference sequence still shows a thin bed of predominantly sandy dolomite with abundant clasts at the base (depth 1700 m) of this formation (Gritsik, 1969). However, in borehole sequences located further to the south, the Manykai Formation is completely devoid of its initial lithological features (Fig. 2), and it is a common practice to distinguish here some formations of inner areas of the Siberian platform in this stratigraphic interval (*Resheniya...*, 1989). The lower part of the Emyaksy Formation, which bears the Tommotian SSF, also loses its variegated color here and is substituted by the Bilir Formation composed of gray limestones and dolomites with marl and mudstone interbeds (Fig. 3).

The correlation of Vendian-Cambrian succession of the Anabar massif and those recovered by boreholes K-203 and K-204 with the stratigraphic scheme of inner areas, which is calibrated on the basis of the borehole section Srednyaya Botuoba 10 (*Resheniya...*, 1989), is mainly based on the exploration logging data.

Figure 2 presents our version of the successive correlation of sections beginning from borehole K-203 in the north to the Srednyaya Botuoba 10 in the south. The relatively persistent Staraya Rechka Formation of rhithic structure can be subdivided into members correlative along the lateral profile. The overall thickness of the Vendian deposits tends to increase southeastward from 300 m (boreholes K-203 and K-204) up to 600 m (borehole Eiik 343-0). The same trend is recorded further to the south: their thickness is 900 m in the borehore Srednyaya Markha and 2250–1050 m in the borehole Verkhnie Vilyuchany 36 (*Resheniya...*, 1989). The thickness decreases again away from the axis of the depression, toward the Nepa-Votuoba Uplift (the right part of Fig. 2). It is characteristic that the thickness reduction is mainly at the expense of the pre-Byuk deposits (Fig. 1). The Byuk and overlying formations of the Upper Vendian of the inner part of the Siberian platform, as well as all Upper Vendian subdivisions of the whole platform, are consistent in thickness and composition.

During the Vendian-Cambrian time (the Manykai and basal Emyaksy formations), a characteristic Early Cambrian facial zonation, corresponding to an open sea in the northeast and an evaporite basin in the southwest, began to form in the Siberian platform. A relatively elevated area separating them (the transitional or Anabar-Sinyaya facies zone) gradually migrated in the northeastern direction (Khomentovskii and Karlova, 1994). In its lithology, the larger upper part of the Emyaksy Formation (borehole Markha-1p) is more similar to the Pestrotsvetnaya Formation of the Anabar-Sinyaya facies zone than to the Emyaksy Formation of the northeastern Anabar region, which was dominated by the open sea environments, as is evident from the widespread algal-archaeothyathid bioherms and specific assemblage of fossils. Therefore, the dashed lines in Fig. 1 show a medial position of the transitional zone. The most significant facial changes, particularly in the Manykai and Emyaksy formations, took place near boundaries of the facies zones.

Although the facial changes hamper correlation based on logging data, our correlation scheme of the considered and deep southern boreholes (Fig. 2) is substantiated by the following reliable facts: (1) the base of the Manykai Formation is correlative with the base of the Kudulakh Formation; (2) the larger lower part of the Yuryakh Formation corresponds to the topmost Manykai and basal Emyaksy beds; and (3) the Bilir Formation is of the same age as the younger beds of the Emyaksy Formation. The suggested correlation is supported by paleontological evidence. In borehole K-203, small-shelled fossils (*Anabarites trisulcatus* Miss. and *Nevadotubulus* sp.) appear first in the depth interval of 958–929 m. These forms are typical of the *Anabarites trisulcatus* Zone. The interval of 901–909 m yielded the more diverse assemblage of *A. trisulcatus* Miss., *A. orbicularis* Bok., *A. grandis* Val., *Angusteochrea* cf. *lata* Val., *Tiksitheca* sp., and hyolithes of the Circothecidae family. This assemblage is typical of the *Purella antique* Zone. In the interval of 866.6–873 m, the zonal assemblage also includes *Ladatheca* sp., *Purella antiqua* (Ab.), and *Platisolenites* sp. Species in the interval of 862–863 m are represented by *Curtitheca korobovi* (Miss.), *Turcutheca* cf. *crasseocochlia* (Sys.), *Loculitheca* sp., *Purella antiqua* (Ab.), *Hyolithellus tenuis* Miss., *H. grandis* Miss., *H. vladimirovae* Miss., *A. trisulcatus* Miss., *A. kelleri* Miss., *A. signatus* Mamb., *Tiksitheca licis* Miss., *Fomitchella acinaciformis* Miss., and others. Along with forms that are reported from older beds, this list includes taxa typical of the *N. sunnaginicus* Zone of the Tommotian Stage. In the interval of 810–828 m, the transitional forms are associated with the first representatives of the *D. regularis–Burites* sp. zones.

In borehole K-204, representatives of small-shelled fossils of the *A. trisulcatus* Zone (*Anabarites* sp. and *Cambrotubulus* sp.) appear first in the lowermost beds of the Manykai Formation (depth 973 m). These fossils successively associate upsection with *A. trisulcatus* Miss. (968.5 m) and *A. grandis* Miss. (935 m). Above the depth level of 908.5 m, there appear forms typical of the *P. antiqua* Zone: *A. signatus* Mamb., *A. kelleri* Miss., and *A. valkovi* Fed. Higher in the section, this assemblage coexists with siphogonuchitids (904.5 m) and *Purella antiqua* (Ab.) (898 m). Forms typical of the *N. sunnaginicus* Zone, such as *Turcutheca crasseocochlia* (Sys.), *Curtitheca korobovi* (Miss.), *Allatheca* sp., *Aldanella crassa* Miss., and *A. attleborensis* (Shal. et Foerste), were found in the interval of 887–892 m only. In the interval of 957–872 m, species known from the *N. sunnaginicus* Zone grow in number: *Nikatheca* sp., *Mellapegma indecora* (Miss.), *Aldanella rozanovi*

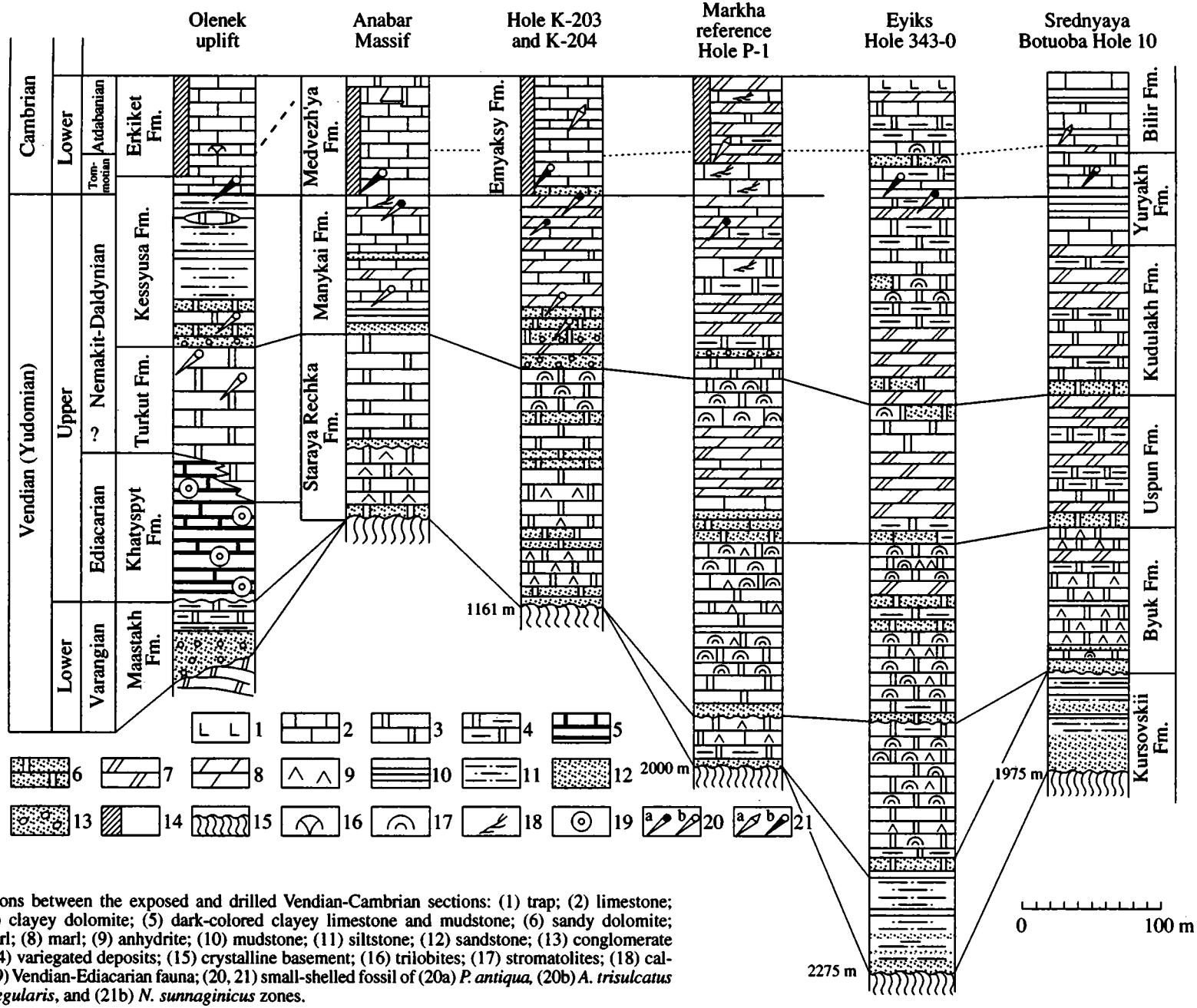


Fig. 3. Correlations between the exposed and drilled Vendian-Cambrian sections: (1) trap; (2) limestone; (3) dolomite; (4) clayey dolomite; (5) dark-colored clayey limestone and mudstone; (6) sandy dolomite; (7) dolomitic marl; (8) marl; (9) anhydrite; (10) mudstone; (11) siltstone; (12) sandstone; (13) conglomerate and gritstone; (14) variegated deposits; (15) crystalline basement; (16) trilobites; (17) stromatolites; (18) calcareous algae; (19) Vendian-Ediacarian fauna; (20, 21) small-shelled fossil of (20a) *P. antiqua*, (20b) *A. trisulcatus* Zone, (21a) *D. regularis*, and (21b) *N. sunnaginicus* zones.

Miss., *Fomitchella infundibuliformis* Miss., and *F. acinaciformis* (Miss.). Moreover, these forms associate here with *Gracilitheca* sp. and *Eonovitatus* sp., which are typical of the upper beds of the *D. regularis* Zone and pass into the Atdabanian Stage.

Thus, in the well-correlated sequences of holes K-203 and K-204, fossils of *A. trisulcatus* Zone begin to appear in the lower beds of the Manykai Formation. They characterize the interval of 65 m in total thickness (Hole K-204), which corresponds in our correlation scheme (Fig. 2) to the main part of the Kudulakh Formation, although fossils have not yet been found in its lower part 40 m thick. The 46-m-thick interval contains here fossils of the *P. antiqua* Zone (Hole K-203) corresponding to the lower pre-Tommotian part of the Yuryakh Formation (Fig. 2). Finally, the first occurrence levels of fossils of the *N. sunnaginicus* Zone and *D. regularis* Zone (Hole K-204) are separated by an interval of 25 m, where 10 m are unfossiliferous. Therefore, the *N. sunnaginicus* Zone may be much reduced in the range.

The presented paleontological characteristics of the Vendian-Cambrian sections revealed by boreholes provide additional criteria for their correlation with both the sequences of the southern deep holes and the outcrops of the relevant strata on the Anabar and Olenek uplifts. In the Markha reference hole, the beds corresponding to the *N. sunnaginicus* Zone are in the interval between the occurrence levels of *Oelandiella,* typical of the *P. antiqua* Zone (1610 m), and *Notabilites* sp., typical of the *D. regularis* Zone (1556 m). Almost the entire interval from the base of the Manykai Formation to the occurrence level of *Oelandiella* bears the calcareous algae *Korilophyton* and *Renalcis jakuticus* (Kolosov, 1977).

The Eiik Hole 343-0 shows the more precise position of the lower boundary of the *N. sunnaginicus* Zone. *Cambrotubulous* sp., *Conotheca* sp., and *Platysolenites* sp. (the *P. antiqua* Zone) occur here in the interval of 1732–1740 m, and the interval of 1720–1728 m bears the very representative assemblage of the *N. sunnaginicus* Zone: *Allatheca* sp., *Ladatheca annae* (Sys.), *Spinulitheca* sp., *Barskovia hemisymmetrica* Gol., *Securiconus consulatus* Miss., *Anabarella* sp., *Obtusoconus* sp., *Maikchanella* sp., *Anabarites kelleri* Miss., *A. signatus* Mamb., *Siphogonuchites triangularis* Qian., *Lopochites latazonalis* Qian., *Halkieria amorphe* (Mesh.), *H.* cf. *saociforma* (Mesh.), and *Watsonella sibirica* (Miss.) in association with fragments of unidentifiable archaeocyathids. Some Tommotian SSF were found only in the Bilir Formation and, probably, in younger deposits recovered by the southern deep holes.

Thus, the paleontological records suggest that the lower boundary of the *N. sunnaginicus* Zone can serve as a good marker for correlation in the northern interiors of the Siberian platform. It is quite natural, because the SSF assemblage abruptly becomes very diverse all over Siberia beginning from this level.

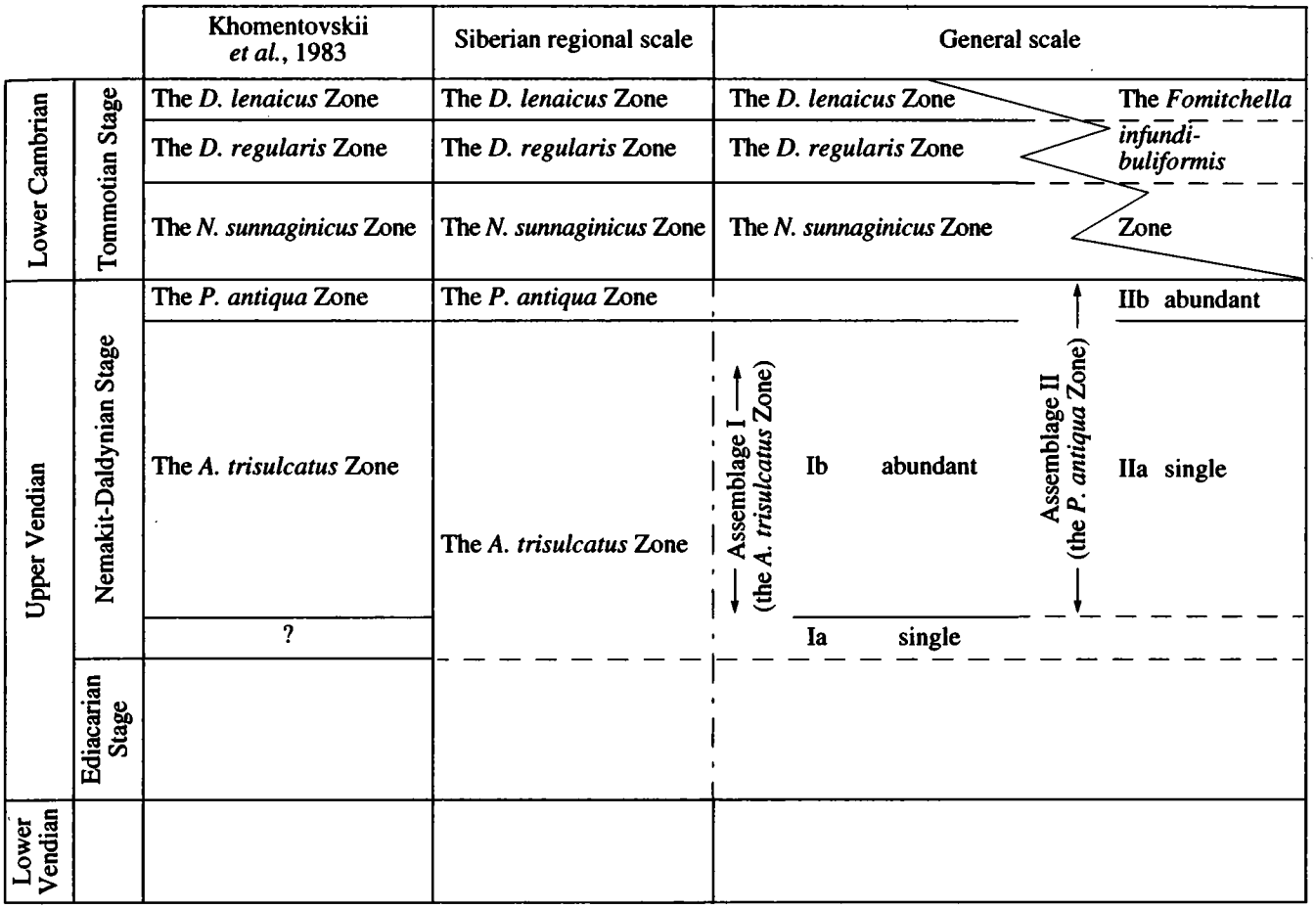
There is some difficulty in establishing stratigraphic ranges of SSF assemblages, because each taxon appearing in all the mentioned zones is also found in younger deposits, frequently at a very high level. Therefore, the only criterium of substantiation of zonal boundaries is the first appearance level of new forms characteristic of certain zones. The zonal SSF assemblages of Siberia were described in earlier publications (Khomentovsky and Karlova, 1993; and others). Below are summary lists of zonal forms found in the outcrops of the Anabar and Olenek uplifts (Fig. 3).

The *A. trisulcatus* Zone: *Cambrotubulus decurvatus* Miss., *Anabarites trisulcatus* Miss., *A. tristichus* Miss., *A. ternarius* Miss., *A. compositus* Miss., *Protohertzina anabarica* Miss., *P. unguliformis* Miss., *Hertzina pensa* Did., and *Sabellidites cambriensis* Jan.

The *P. antiqua* Zone: hyolithes *Ladatheca annae* Miss., *L. dorsocava* Miss., *Turoutheca* sp., and *Loculitheca* sp.; molluscs *Purella antiqua* (Ab.), *P. cristata* Miss., *Latouchella korobkovi* (Vost.), *Anabarella plana* Vost., *Salanyella costulata* Miss., *Barskovia* sp., *Granoconus trematus* Yu, and *Yantzeconus priscus* Yu; problematics *Lopochites* sp., *Siphogonuchites* sp., *Halkieria sacciforma* (Mesh.), *H. uncostata* Qian et Yin, *Fomitchella* sp., *Hyolithellus tschuscunensis* Val., and *H.* sp.; angustiocreids *Anabarites trisulcatus* Miss., *A. tristichus* Miss., *A. tripartitus* Miss., *A. ternarius* Miss., *A. tricarinatus* Miss., *A. signatus* Mamb., *A. modestus* Bokova, *A. conicus* Miss., *Longiochrea rugosa* Val. et Sys., *Lobiochrea natella* Val. et Sys., *Angustiochrea lata* Val. et Sys., *Aculeochrea ornata* Val. et Sys., *Tiksitheca licis* Miss., *T. korobkovi* Miss., and *Cambrotubulus decurvatus* Miss.

The *N. sunnaginicus* Zone: hyolithes *Turoutheca crasseocochlia* (Sys.), *T. cotuiensis* (Sys.), *Allatheca concinna* Miss., *Ovalitheca mongolica* Sys., *Spinulitheca billingsi* (Sys.), *S. rotunda* Val., *Nikatheca kengedeica* Val., *N. digna* Val., *Kugdatheca voluta* Miss., *Kotuitheca curta* Miss., *Curtitheca* sp., *Ladatheca dorsocava* Miss., *Circotheca obesa* Qian, and *Egdetheca* sp.; molluscs *Aldanella attleborensis* Shal. et Foerst., *A. rozanovi* Miss., *A. crassa* Miss., *Bemella jacutica* Miss., *Igorella ungulata* Miss., *Latouchella memorabilis* Miss., *Anabarella plana* Vost., *Barskovia hemisymmetria* Gol., *Pollicina* sp., *Salanyella* sp., and *Ilsanella* sp.; angustiocreids *Anabarites kelleri* Miss., *Angusteochrea lata* Val. et Sys., *Yacuteochrea lenta* Bokova et Val., *Selindeochrea teota* Val., and *Anabarithellus hexasulcatus* Miss.; archaeostracans *Watsonella sibirica* Miss.; problematics *Triplicatella disdoma* Conway Morris, *Emarginoconus mirus* Yu, *Blastulospongia monotalamus* J. Pickettet et P.P. Jell, *Fomitchella infundibuliformis* Miss., *F. acinaciformis* Miss., *Tommotia kozlowskii* (Miss.), *T. admiranda* Miss., *Prooneotodus* sp., *Phabdochites* sp., *Hyolithellus vladimirovae*

The Vendian-Cambrian biostratigraphy based on SSF



Miss., *H. tenuis* Miss., and *Paragioborilus subglobosus* (He).

The comparative analysis of these lists leaves no doubt that in the Vendian-Cambrian stratigraphic scheme for the inner areas of the Siberian platform the *A. trisulcatus* Zone corresponds to almost the entire Kudulakh Formation; the *P. antiqua* Zone, to the larger lower part of the Yuryakh Formation; and the *N. sunnaginicus* Zone, to the uppermost beds of the Yuryakh Formation and a part of the Bilir Formation (Figs. 2, 3).

It should be noted that the SSF zonation for Siberia is somewhat idealized, and this should be taken into consideration in geological practice. As mentioned above, the forms that first appear at the base of the Tommotian Stage in the transitional facies region very gradually populate sections in the eastern facies region, and their distribution may last up to the Atdabanian Stage in some sections (Khomentovskii and Karlova, 1992, 1994). Therefore, it is difficult to establish the position of the lower Tommotian boundary in concrete sections. Detailed investigations make it possible to draw the boundary on the basis of the first appearances of forms of the *N. sunnaginicus* Zone of the transitional region and associated taxa of the eastern region. In addition, this level also marks the appearance of some archaeocyathiatids and other taxa. However, geological surveys with inadequate paleontological sampling may result in the situation where the enormous interval equal to the entire Tommotian Stage will be characterized only by forms of the *N. sunnaginicus* Zone. Therefore, we suggested the term "*Fomitchella infundibuliformis* Zone" for deposits with such a stratigraphic range and paleontological characteristics (Khomentovskii and Karlova, 1992).

The problem of the two lower *P. antiqua* and *A. trisulcatus* zones is more complicated. In the majority of sections, their boundaries are readily recognizable. The lower boundary of the *A. trisulcatus* Zone is marked by the appearance of respective problematics at the base of Member III of the Ust'-Yudoma Formation in the south of the Siberian platform and at the bases of the Manykai and Kessyusa formations in the north (Khomentovskii and Karlova, 1992; Fedorov and Shishkin, 1984; and others). As mentioned above, the zonal forms appear approximately at the same level in the northern inner areas of the Siberian platform. However, other sections display their occurrence in older beds. For example, *Anabarites* sp. and *Cambrotubulus* sp. were found within the upper 30 m of the Turkut Formation of the

Olenek Uplift, and *Anabarites trisulcatus* Miss., *A. valkovi* Fed., *Protohertzina anabarica* Miss., *P. unguliformis* Miss., and *Olivooides multisulcatus* Qian were reported from the base of Member II of the Ust'-Yudoma Formation in the Uchur River basin, the southern Siberian platform (Khomentovskii and Karlova, 1992).

Fossils of the *P. antiqua* Zone appear at the base of Member IV of the Ust'-Yudoma Formation in many sections at the south of the Siberian platform and in its analogues in the north (Khomentovskii and Karlova, 1992). However, *Lophotheca socialis* Val., *Archaeospira* sp., *Oelandiella* sp., *Purella oristata* Miss., *Siphogonuchites* aff. *triangularis* Qian, and *Tiksitheca korobovi* Miss., typical of the *P. antiqua* Zone, were identified in the lower beds of Member III of the Ust'-Yudoma Formation in Konus Mountain of the southern Siberian platform. In northern areas of the Siberian Platform, along the Olenek River basin, at the mouth of the Ochchugui Kersyuke River, an exposure of the basal Kessyusa Formation yielded *Tiksitheca licis* Miss., *Anabarites kelleri* Miss., and unidentifiable fragments of large tubular forms, which were never recorded in the *A. trisulcatus* Zone. Finally, *Anabarella*? sp. was recovered in borehole N 2 in the lower stream of the Bol'shaya Kyonamka River, in the lower member of the Manykai Formation.

These occurrences formally suggest that the lower boundary of the *P. antiqua* Zone should be placed at the base of beds with fossils of the *A. trisulcatus* Zone, and the basal boundary of the latter zone should be lowered accordingly. However, in this case we would lose an excellent tool for subdivision and correlation of the Vendian-Cambrian deposits over vast regions of Siberia and adjacent territories. Therefore, we suggest retaining, until new information on the regional Siberian scale is available, the present scope of the *P. antiqua* Zone and enlarging that of the *A. trisulcatus* Zone at the expense of beds ranging from the top of the Ediacarian Stage to the base of the Manykai Formation. This is also justified by the occurrence of a complete assemblage of the *trisulcatus* Zone in these beds.

However, it cannot be excluded that, outside the Siberian platform, relatively rich SSF assemblages occur at levels that are marked in Siberia by extremely scarce forms. Therefore, it should be admitted that, within the stratigraphic ranges of the *trisulcatus* and *antiqua* zonal assemblages of the general Vendian-Cambrian biostratigraphic scale, some beds with single representatives of the corresponding assemblages (Ia and IIa in table) and those with abundant forms (Ib and IIb) can be distinguished as overlapping each other in time. If this overlap should appear to be due to local facial or taphonomic environments (Khomentovskii and Karlova, 1994), no significant changes should be made in the biostratigraphic scale. We assume that finds of the *antiqua* zonal species will be extremely rare in the *trisulcatus* Zone. However, if they are abundant, then the interval of overlapping of subassemblages Ib and IIa should be distinguished as an independent subdivision. In this case, the *trisulcatus* Zone would be equal in range to the beds with Subassemblage Ia, because only they would match the initial paleontological characteristics of this zone.

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Permian Stratigraphy and Rugosa Assemblages from the Southeastern Pamir

T. G. Il'ina

Paleontological Institute, Russian Academy of Sciences, Profsoyuznaya ul. 123, Moscow, 117868 Russia
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Abstract—The Rugosa zonation and assemblages from formations and stages of the Bolorian (Lower Permian) to the Dorashamian (Upper Permian) are first defined in the Permian deposits of the southeastern Pamirs. Two new genera (*Timorocarinophyllum* and *Pseudoverbeekiella*) and ten new species are described in detail. Synonymy of several families is revised on the basis of septa microstructure. Correlation with other Tethyan basins is performed.

Key words: *Southeastern Pamirs, stratigraphy, Permian, rugose corals, taxonomy, morphology, microstructure.*

The southeastern Pamirs, with the Permian deposits spanning the Sakmarian to the Dorashamian stages, represents one of a few areas with an almost continuous Permian section. Beginning from the Bolorian Stage, these deposits contain diverse faunal groups, including rugose corals (Table 1). The latter includes 24 genera and 67 species, among which I previously described 2 genera and 12 species (Il'ina, 1970, 1997).[1] Some of my identifications were cited in works by Leven (1967) and Grunt and Dmitriev (1973) and were also used in the explanatory notes to the regional stratigraphic scheme (Leven *et al.*, 1989). The study of corals was conducted on the basis of my own materials, as well as on collections of Leven, Grunt, Dmitriev, D.V. Dronov, S.V. Ruzhentsev, S.V. Rozhnov, and V.P. Novikov (in total, about 1000 specimens).

STRATIGRAPHIC CHARACTERISTICS

The Permian deposits outcrop in the Murgab–Aksu fold zone, stretching in the northwest–southeast direction (Fig. 1). Based on structural features and thickness of deposits, the Murgab–Aksu zone is subdivided into three subzones: central, intermediate, and marginal subzones (Leven, 1967; Leven *et al.*, 1989). Rugose corals occur only in the first two subzones. Subdivision into stages was performed in accordance with the regional Permian scale for Central Asia (Leven *et al.*, 1989) accepted by the Interdepartmental Stratigraphic Committee June 4, 1981. Only some modifications concerning the age of the uppermost Permian deposits were introduced into it according to new data on conodonts (Davydov *et al.*, 1991). The Kutal' Member of the Gana Formation was referred by conodonts to the Dzhulfian Stage, whereas the middle Takhtabulak Subformation was placed in the Dorashamian Stage to obtain agreement with the data on fusulinids. The boundary of the Dzhulfian Stage is drawn at the top of the Karasin Member corresponding to the Midian Stage.

Corals were sampled in the Chakobai, Mamazair-Bulak, Kur-Teke, and Kastanat-Dzhilga sections of the central subzone and in the Dzhamantal, Yuzhnyi Ak-Arkhar, Kutal' 1, Muzdu-Bulak, Ak-Bura, Kur-Istyk, Sulyu-Dzhilga, Shindy, Dunkel'dyk, and Igrimyuz sections of the intermediate subzone. The Igrim'yuz section is transitional to sections of the central subzone. The intermediate subzone differs from the central subzone by a greater thickness of sections containing a great amount of volcanogenic material and siliceous rocks (Fig. 2). Table 2 demonstrates the distribution of rugose corals throughout the sections.

The Bolorian Stage. Deposits of the Bolorian Stage transgressively overlie the age-variable strata of the Tashkazyk Formation, because deposits of the Yakhtashian Stage are missing in the Murgab–Aksu zone.[2] The Kochusu and Shindy formations are defined in the Bolorian Stage (Bolorian Horizon). The Kochusu Formation is subdivided into the lower and upper subformations. In the central subzone (Mamazair-Bulak, Kastanat-Dzhilga, and Chakobai sections) and in sections transitional from this to the intermediate subzone (Igrim'yuz), the lower subformation is composed of organogenic-detrital limestones (1.5–3 m) with rugose and tabulate corals and ammonoid, brachiopod, crinoid, and conodont fossils. Corals are represented by a rich assemblage of abundant and diverse, mainly solitary, Rugosa: *Pentaphyllum permicum, P. gerthi, Ufimia multitabulata, U. variabilis, U. crassisepta, U. super-*

[1] Only brief diagnoses of new genera and species were given.

[2] Deposits of the Sakmarian Stage devoid of rugose corals are not considered here.

Table 1. Correlation of the Permian Rugosa assemblages with reference faunal groups

System	Series	Stage	Formation	Subformation	Rugosa	Foraminifers (Leven *et al.*, 1967)	Conodonts (Davydov *et al.*, 1991)	Ammonoids (Chediya *et al.*, 1986; Leonova *et al.*, 1989)
Permian	Upper	Dorashamian	Takhtabulak	middle	*Pseudohuangia collucata*	*Calaniella parva*	*Gondolella subcarinata*	
					Waagenophyllum (W.) tachtabulasicum	*C. lepida*	*G. planta*	
					W. (W.) crassiseptum	*C. cylindrica*		
					W. (W.) ramosum	*Abadehella coniformis*		
					W. (Huayunophyllum) pamiricum	*Reichelina cribroseptata*		
		Dzhulfian		lower	*Asserculinia prima*	*Stafella zisongzhengensis*	*G. bitteri*	
					Pseudohuangia pyzhjanivi	*Reichelina pulchra*	*G. rosenkrantzi*	
			Gana		*W. (W.) kueichowense*	*Codonofusiella curtekensis*	*G. intermedia*	
					W. (W.) virgalense	*Paradunbarula pamirica*	*Diplognathodus oertlii*	
					W. (Huayunophyllum) pamiricum	*Calaniella parva*	*Iranognathus taraz*	
		Midian			*Waagenophyllum (W.) kueichowense*	*Neoschwagerina margaritae*	*G. bitteri*	
					Pavastephyllum (T.) carinatiseptatum	*N. minoensis*	*G. rosenkrantzi*	
					P. (T.) carcinophylloides	*Afghanella sumatrinaeformis*	*Diplognathodus oertlii*	
					Iranopyllum (I.) splendens	*Yabeina archaica*	*Sweetognatus* sp.	
					I. (Laophyllum) aff. *pongouaense*	*Verbeekiina verbeeki*		
		Murgabian			*Waagenophyllum (W.) kueichowense*	*Neoschwagrina craticulifera*	*G.* aff. *rosenkrantzi*	
					Pavastephyllum (T.) carinatiseptatum	*Pseudoliolina ozawai*	*Diplognathodus oertlii*	
					Iranopyllum (I.) amygdalophylloideum	*Cancellina dutkevitchi*		
					I. (Laophyllum) igrimjusicum	*Armenina karinae*		
					Lonsdaleiastraea tipica	*Neoschwagerina simplex*		
		Kubergandian	Kubergandy		*Praewentzelella magnifica*	*Cancellina cutalensis*		*Bamyaniceras kubergandense*
					Timorocarinophyllum zolkinae	*Cancellina primigena*		*Paraceltites edelsteini*
					Wannerophyllum exedens	*Armenina pamirensis*		*Popanoceras pamiricum*
					Kabakovitchiella duplex	*Armenina salgirica*		*P. kubergandense*
					Duplophyllum tenuiseptatum	*Misellina (M.) ovalis*		*Propinacoceras karpinskyi*
					Pavastephyllum (T.) simplex			*Stacheoceras discoidale*
	Lower	Bolorian				*Misellina (M.) claudiae*		
			Shindy		*Yatsengia pachytheca*	*Darvasites contractus*		*Daraelites pamiricus*
					Euryphyllum cainodon	*Misellina aliciae*		*Medlicottia tenuis*
					Wannerophyllum exedens	*Pseudofusulina* ex gr. *kraffti*		*Propinacoceras toumanskayae*

Table 1. (Contd.)

System	Series	Stage	Formation	Subformation	Rugosa	Foraminifers (Leven *et al.*, 1989)	Conodonts (Davydov *et al.*, 1991)	Ammonoids (Chediya *et al.*, 1986; Leonova *et al.*, 1989)
Permian	Lower	Bolorian	Kochusu	upper	*Pentaphyllum gerthi* *Ufimia variabilis* *Timorocarinophyllum zolkinae* *Verbeekiella arboricalis*	*Monodiexodina shiptoni*	*G. idahoensis* *Neostreptognathodus leonovae* *Diplognathodus nassichuki*	*Cardiella parva* *Parapronorites rectus* *Metalegoceras toumanskayae* *Crimites shyndensis*
				lower	*Pentaphyllum permicum* *Ufimia persimmetrica* *Timorocarinophyllum zolkinae* *Lophophyllidium spinosum* *Wannerophyllum cristatum* *Pseudoverbeekiella variabilis* *Wannerophyllum excedens*		*Neostreptognathodus sulcoplicatus* *Gondolella idahoensis* *Hindeodus minutus* *Duplognathodus nassichuki* *Gnatodus artinskiensis*	

ans, U. alternata, U. persimmetrica, Calophillum dobrolyubovae, Timorocarinophyllum zolkinae, Pseudoverbeekiella variabilis, Verbeekiella gerthi, V. arboricalis, Lophophyllidium (Lophbillidium) spinosum, L. (Lophbillidium) martini, Wannerophyllum cristatum, W. excedens, W. aff. *tubulosum, Cyathaxonia minima, C. orientalis, Amlexocarina jonkeri, A. socialis, Kabakovitchiella* aff. *duplex, Asserculina transcarinata, Euryphyllum reidi, Basleophyllum incertum, B. pusillum,* and *Pseudohuangia stocklini.* Deposits of the lower Kochusu Subformation in the central subzone are represented only by these basal limestones. In the intermediate subzone (Kur-Istyk, Shindy, Yuzhnyi Ak-Arkhar, and Ak-Bura sections), the base of the subformation is marked by the gray clayey limestone bed (1–3 m) with the impoverished coral assemblage consisting of five species: *Timorocarinophyllum zolkinae, Cyathaxonia orientalis, Pseudoverbeekiella variabilis, Lophophyllidium (Lophbillidium) spinosum,* and *Ufimia crassisepta.* Calacareous aleuropelites (10–15 m) with limestone interbeds occurring upsection yielded rare solitary rugose corals *Timorocarinophyllum zolkinae.*

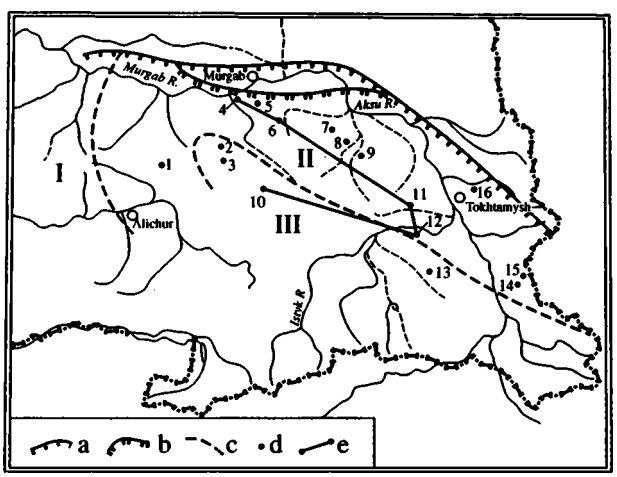

Fig. 1. Localities of sections and Rugosa finds. Structural–facial subzones of the Murgab-Aksu fold zone: (I) Marginal; (II) Intermediate; (III) Central. Sections and fauna localities: (1) Yuzhnyi Ak-Arkhar; (2) Chakobai; (3) Mamazair-Bulak; (4) Dzhamantal; (5) Kutal'; (6) Muzdu-Bulak; (7) Shur-Bulak; (8) Sulyu-Dzhilga; (9) Ak-Bura; (10) Kur-Teke; (11) Kur-Istyk; (12) Igrim'yuz; (13) Kastanat-Dzhilga; (14) left side of the Shindy River; (15) upper stream of the Shindy River; (16) Dunkel'dyk; (a) the northern boundary of geological structures of the southeastern Pamirs (Pshart thrust); (b) Murgab thrust; (c) boundary between structural–facial subzones; (d) studied sections and localities; (e) correlation lines.

The upper Kochusu Subformation of the central subzone (the Mamazair section) consists of silty limestones (5–6 m), whereas in other sections (Chakobai, Kastanat-Dzhilga) it is composed of dark green tuffs and limestones with a tuffaceous admixture and detrital limestone interbeds. These rocks bear corals: *Lophophyllidium (Lophbillidium) spinosum, Timorocarinophyllum zolkinae, Pseudoverbeekiella variabilis, Wannerophyllum cristatum, W. excedens,* and *Ufimia variabilis,* as well as rare ammonites. In the intermediate subzone, the respective deposits are represented by a member of dark gray claystones and clayey siltstones with limy sandstone interbeds (15–35 m) containing rare rugose corals *Timorocarinophyllum zolkinae* and *Ufimia variabilis.*

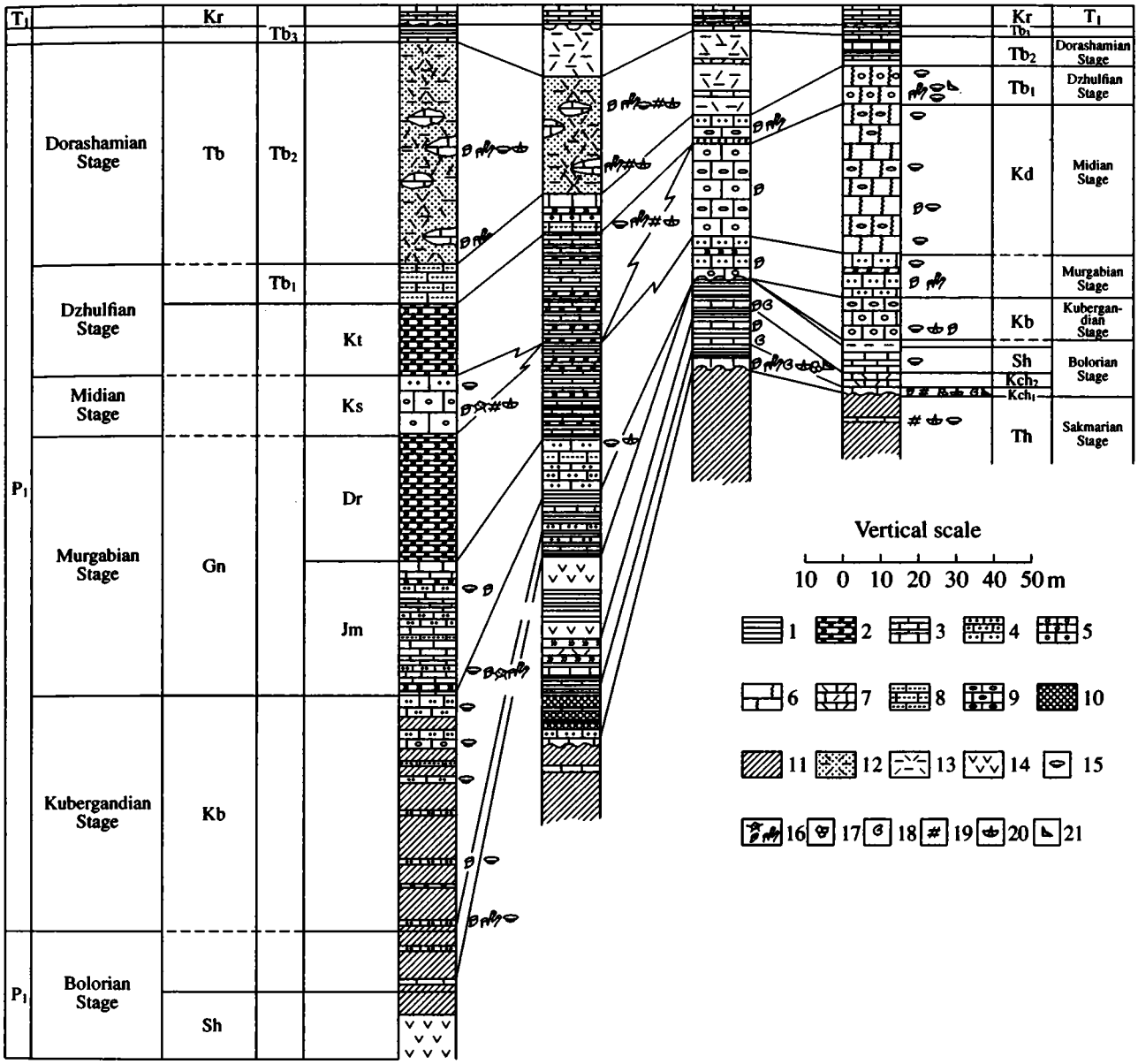

Fig. 2. Correlation of Permian sections in the Murgab-Aksu zone along the line Kutal'–Kur-Istyk–Igrim'yuz (intermediate subzone)–Kur-Teke (central subzone). Sections are compiled using the profile of Grunt and Dmitriev (1973): (Th) Tashkyzak Formation; (Kch) Kochusu Formation; (Sh) Shindy Formation; (Kb) Kubergandy Formation; (Gn) Gana Formation with (Jm) Dzhamantal, (Dr) Deire, (Ks) Karasin, and (Kt) Kutal' members; (Kd) Kastanat-Dzhilga Formation; (Tb) Takhtabulak Formation; (Kr) Kobrigen Formation of the Lower Triassic; (1) shale; (2) chert; (3) platy pelitomorphic limestone; (4) detrital limestone; (5) conglomerate-like limestone; (6) massive hydroid–algal reef limestone; (7) limestone with the tuffaceous admixture; (8) sandy limestone; (9) limestone with cherty inclusions; (10) calcareous siltstone; (11) clayey siltstone and argillite; (12) tuffaceous sandstone; (13) tuffaceous argillite; (14) basic effusives; (15) fusulinids; (16) Rugosa; (17) Tabulata; (18) ammonoids; (19) bryozoans; (20) brachiopods; (21) conodonts.

Thin (4–8 m) fusulinid and detrital limestone beds with single corals *Euryphyllum cainodon, Yatsengia pachyteca, Amplexocarinia* aff. *subtilis*, and *Kabakovitcheiella* aff. *arcuata* are referred in the central subzone to the Shindy Formation. In the intermediate subzone, the formation is composed of greenish and black shales; thick lava and basic tuff sequences with limestone lenses; and, locally (Shindy Fm.), lava flows with spheroid parting (up to 70–100 m), the calcareous matrix of which contains rare brachiopods, ammonites, conodonts, and rugose corals *Wannerophyllum excedens.*

The Bolorian Stage also comprises the lower part of the Kubergandy Formation, which is composed of greenish siltstones and claystones in the intermediate subzone (the *Misellina claudiae* Zone), which is devoid of rugose corals and ammonoids. No outcrops of these deposits are observed in the central subzone.

Table 2. Distribution of Rugosa in the southeastern Pamirs and their correlation with corals from other regions

Rugosa assemblages	N	Rugosa species	Lower Permian			Upper Permian				
			Bolorian Stage			Kubergandian Stage	Murgabian Stage	Midian Stage	Dzhulfian Stage	Dorashamian Stage
			Kochusu Formation		Shindy Formation	Kubergandy Formation	Gana Formation		Takhtabulak Formation	
			lower subformation	upper subformation					lower subformation	middle subformation
1	2	3	4	5	6	7	8	9	10	11
Waagenophyllum (W.)–W. (Huayunophyllum) Zone ((Takhtabulak assemblage)	1	*Kabakovitchiella arcuata* (Schouppe et Stacul, 1959)								
	2	*Pseudohuangia collucata* Iljina, 1997								
	3	*P. pyzhjanovi* Iljina, 1997								
	4	*Waagenophyllum (W.) tachtabulasicum* Iljina, 1997								
	5	*W. (W.) crassiseptum* Iljina, 1997								
	6	*W. (W.) ramosum* Iljina, 1997								
	7	*W. (W.) virgalense* (Waagen et Wentzel, 1886)								
	8	*W. (W.)* aff. *virgalense* (Waagen et Wentzel, 1886)								
	9	*W. (Huayunophyllum) pamiricum* Iljina, 1997								
Pavastephyllum–Iranophyllum Zone (Gana assemblage)	10	*Asserculinia prima* Schouppe et Stacul, 1959								
	11	*Amplexocarinia abichi* (Waagen et Wentzel, 1886)								
	12	*Epiphanophyllum sinuosum* Iljina, 1970								
	13	*Waagenophyllum (W.) kueichowense* Huang, 1932								
	14	*Pavastephyllum (Tomasiphyllum) carinatiseptatum* Zhao et Wu, 1986								
	15	*P. (Tomasiphyllum) carcinophylloides* (Douglas, 1936)								
	16	*Iranophyllum (I.) splendens* Douglas, 1936								
	17	*I. (I.) amygdalophylloideum* (Huang, 1932)								
	18	*I. (Laophyllum?) igrimjusicem* Iljuna, 1997								
	19	*I. (Laophyllum)* aff. *pongouaense* (Mansuy, 1912)								
	20	*Lonsdaleiastraea tipica* Gerth, 1938								
	21	*L.* aff. *yamanbaensis* Minato, 1949								
	22	*Ufimia* aff. *persimmetrica* Schindewolf, 1942								
	23	*U. beyrichi* (Rothpletz, 1892)								
	24	*U. kobayashi* Schindewolf, 1942								
	25	*U.* aff. *alternata* (Huang, 1932)								
	26	*Cyathaxonia* sp.								
	27	*C. khmeriana* Fromentel, 1961								
Kubergandy assemblage	28	*Cyathocarinia multituberculata* Soshkina, 1928								
	29	*Amplexocarinia heimoi* Heritch, 1936								
	30	*Kabakovitchiella duplex* (Schouppe et Stacul, 1959)								
	31	*Duplophyllum tenuiseptatum* Schouppe et Stacul, 1959								
	32	*Pseudohuangia* aff. *tabellata* (Douglas, 1936)								
	33	*Pavastephyllum (Tomasiphyllum) simplex* (Douglas, 1936)								
	34	*Praewentzelella magnifica* (Douglas, 1936)								
	35	*Yatsengia asiatica* Huang, 1932								

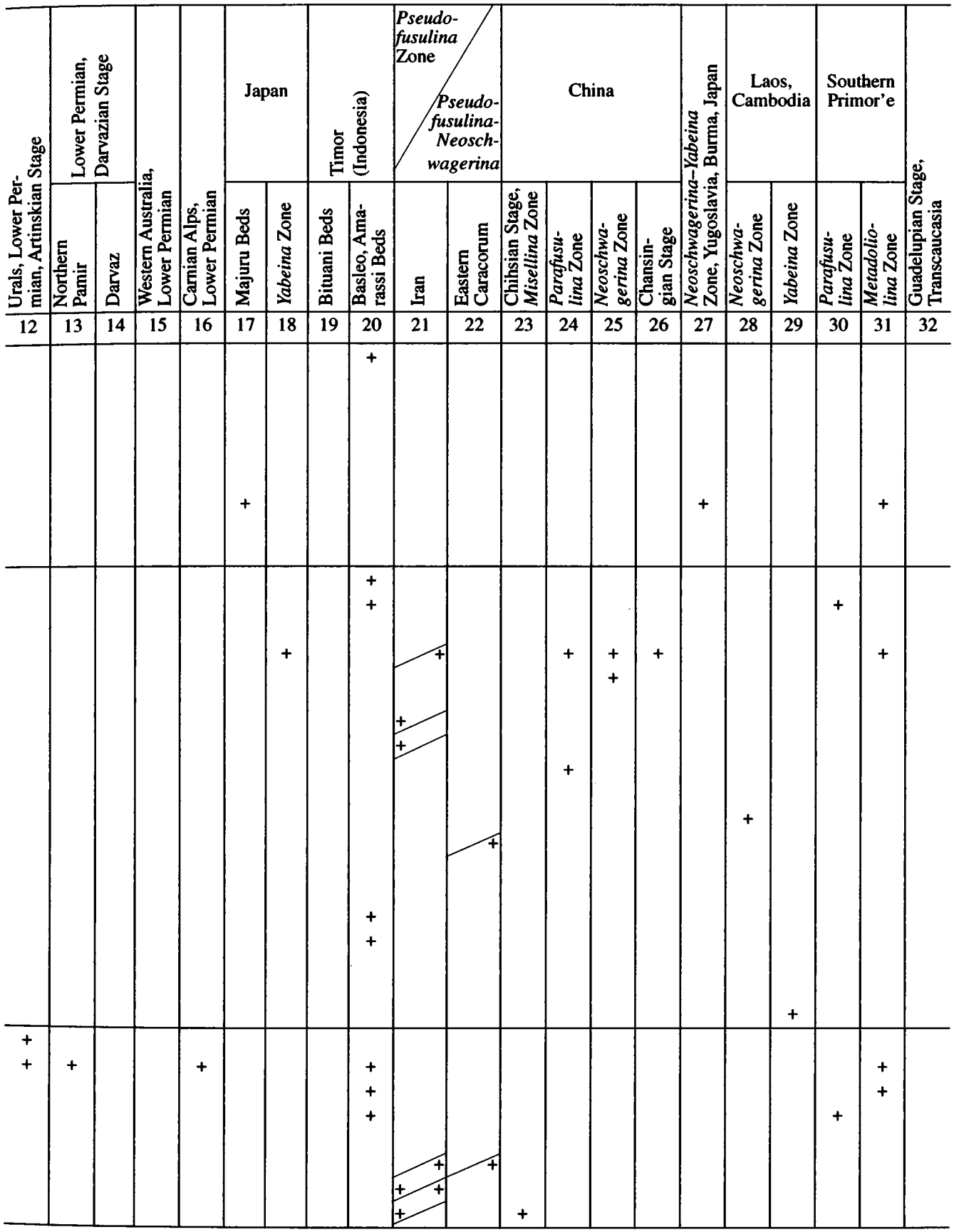
Urals, Lower Permian, Artinskian Stage
Lower Permian, Darvazian Stage
Northern Pamir
Darvaz
Western Australia, Lower Permian
Carnian Alps, Lower Permian
Japan
Majuru Beds
Yabeina Zone
Timor (Indonesia)
Bituani Beds
Basleo, Amarassi Beds
Pseudofusulina Zone
Pseudofusulina-Neoschwagerina
Iran
Eastern Caracorum
China
Chihsian Stage, Misellina Zone
Parafusulina Zone
Neoschwagerina Zone
Chansingian Stage
Neoschwagerina–Yabeina Zone, Yugoslavia, Burma, Japan
Laos, Cambodia
Neoschwagerina Zone
Yabeina Zone
Southern Primor'e
Parafusulina Zone
Metadoliolina Zone
Guadelupian Stage, Transcaucasia
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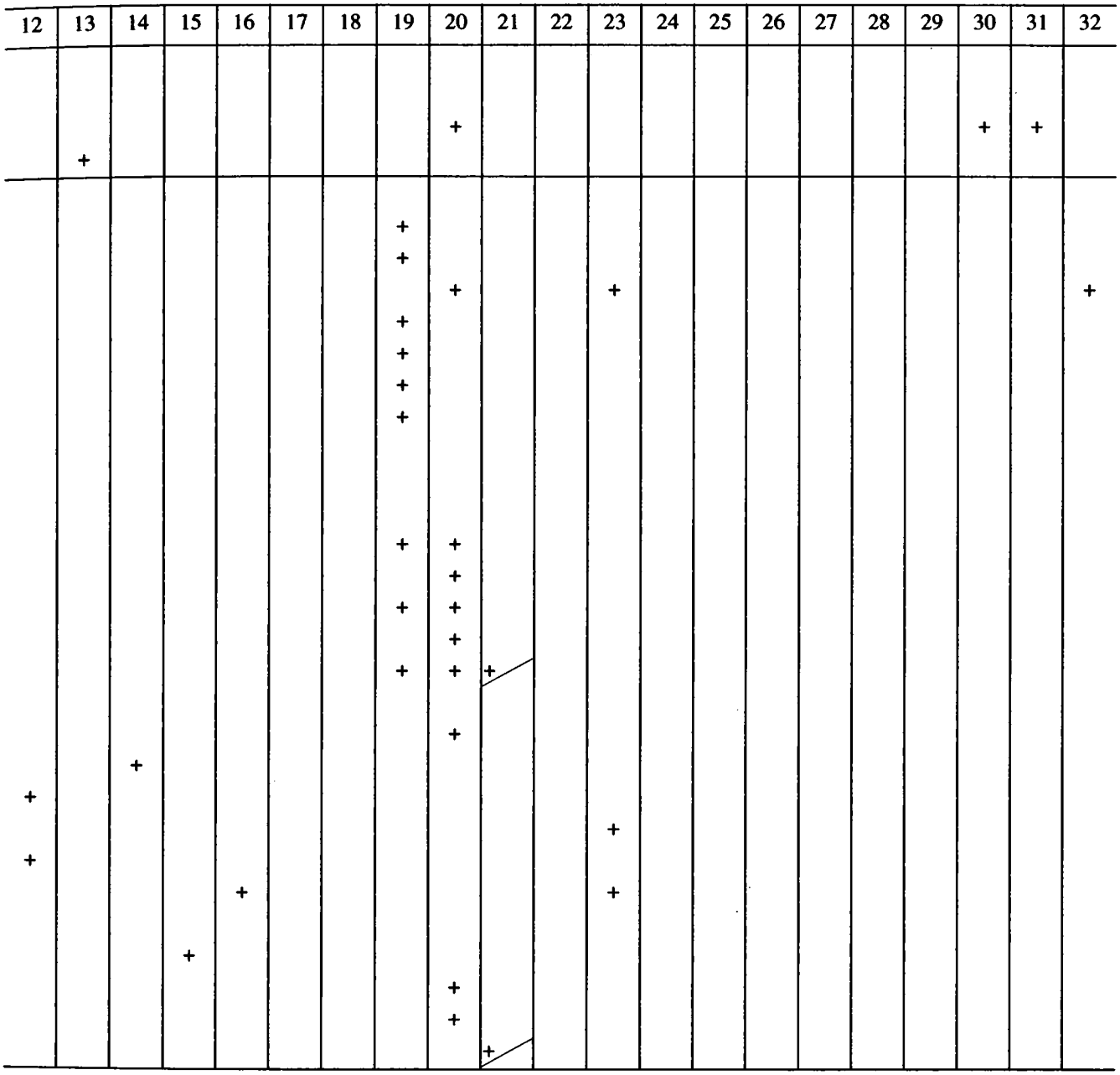
Table 2. (Contd.)

1	2	3	4	5	6	7	8	9	10	11
Shindy assemblage	36	*Amplexocarinia* aff. *subtilis* (Schouppe et Stacul, 1959)			—					
	37	*Kabakovitchiella* aff. *arcuata* (Schouppe et Stacul, 1959)			—					
	38	*Euryphyllum cainodon* (Koker, 1924)			—					
	39	*Yatsengia pachytheca* Pyzhjanov, 1966			—					
Pseudoverbeekiella–Timorcarinophyllum–Ufimia Zone (Kochusu assemblage)	40	*Pentaphyllum permicum* Iljina, 1997	—							
	41	*P. gerthi* (Schindewolf, 1942)	—	—						
	42	*Ufimia multitabulata* Schindewolf, 1940	—							
	43	*U. alternata* (Huang, 1932)	—							
	44	*U. persimmetrica* Schindewolf, 1942	—							
	45	*U. superans* Schindewolf, 1942	—							
	46	*U. crassisepta* (Schindewolf, 1942)	—							
	47	*U. variabilis* (Schindewolf, 1940)	—	—						
	48	*Calophyllum dobrolyubovae* (Iljina, 1970)	—							
	49	*Timorocarinophyllum zolkinae* Iljina, 1997	—			—	—	—		—
	50	*Pseudoverbeekiella variabilis* Iljina, 1997	—	—						
	51	*Verbeekiella gerthi* Schouppe et Stacul, 1955	—							
	52	*V. atboricalis* Schouppe et Stacul, 1955	—							
	53	*Lophophyllidium (Lophbillidium) spinosum* (Martin, 1881)	—	—						
	54	*L. (Lophbillidium) martini* Schouppe et Stacul, 1955	—							
	55	*Wannerophyllum cristatum* (Gerth, 1921)	—	—						
	56	*W.* aff. *tubulosum* (Gerth, 1921)	—							
	57	*W. excedens* Schouppe et Stacul, 1955	—		—	—				
	58	*Cyathaxonia minima* Pyzhjanov, 1966	—							
	59	*C. orientalis* Dobrolyubova, 1940	—							
	60	*Asserculinia transcarinata* Zhao et Wu, 1986	—							
	61	*Amplesocarinia socialis* Soshkina, 1941	—							
	62	*A. jonkeri* (Koker, 1924)	—							
	63	*Kabakovitchiella* aff. *duplex* (Koker, 1924)	—							
	64	*Euryphyllum reidi* Hill, 1938	—							
	65	*Basleophyllum incertum* (Koker, 1924)	—							
	66	*B. pusillum* Schouppe et Stacul, 1959	—							
	67	*Pseudohuangia stocklini* Minato et Kato, 1965	—							

The Kubergandian Stage. Upsection, the Upper Permian deposits of the Kubergandian Stage (Horizon and Formation) occur. In the central subzone, its lower boundary is distinct owing to the lighter coloration of limestones. The largest portion of the Upper Permian deposits in this subzone is represented by hydroid reefs (Kur-Teke and Chakobai sections). These are thick-bedded light gray limestones (8–12 m), locally with chert inclusions. Limestones incorporate fusulinids, brachiopods, gastropods, crinoids, and rugose corals *Timorocarinophyllum zolkinae, Amplexocarinia heimoi,* and *Pseudohuangia* aff. *tabellata.* In the intermediate subzone (Ak-Bura, Kutal', and other sections), the Shindy Formation is overlain by greenish gray psammitic tuffites or by sandy-detrital limestones with the volcanogenic admixture and, locally, with dark gray claystone and tuffaceous sandstone interbeds 12–90 m thick (Kutal' section). Faunal remains are represented by foraminifers, brachiopods, crinoids, rare ammonites, and rugose corals *Timorocarinophyllum zolkinae, Wannerophyllum excedens, Cyathocarinia multituberculata, Kabakovitchiella duplex, Duplophyllum tenuiseptatum, Yatsengia asiatica, Pavastephyllum (P.) simplex, Praewentzelella magnifica,* and *Pseudohuangia* aff. *tabellata.*



Deposits of the Murgabian, Midian, and, partly, Dzhulfian stages are united into the Kastanat-Dzhilga Formation in the central subzone and into the Gana Formation in the intermediate subzone.

The Murgabian Stage. The Murgabian Stage of the central subzone includes the Dzhamantal Member of thick-bedded light gray detrital limestones with cherty interbeds and inclusions and also the lower part of the Kastanat-Dzhilga Formation of biohermal limestones (12–20 m). In the Kur-Teke section, these limestones yielded rugose corals *Waagenophyllum kueichowense, Ufimia* aff. *kobayashi,* and *Pavastephyllum* (*Tomasiphyllum*) *carinatiseptatum.* In the intermediate subzone, the Murgabian Stage consists of two units: the Dzhamantal Member represented by alternating cherty–marly shales and detrital limestones (40–70 m) and the Deire Member of greenish violet thin-platy cherty limestones and silicified marly shales (15–40 m).[3] Brachiopods, bryozoans, and rugose corals occur only in the Dzhamantal Member. Rugose corals are represented by *Timorocarinophyllum zolkinae, Ufimia beir-*

[3] Locally, in the upper part of the section, where the Karasin Member wedges out and is replaced along the strike by the Deire Member, the latter is the Midian in age.

ichi, U. aff. *alternata, Cyathaxonia khmeriana, Iranophyllum (I.) igrimjusicum, I. (I.) amygdalophylloideum, Lonsdaleiastraea typica,* and *Pavastephyllum (Tomasiphyllum) carinatiseptatum.*

The Midian Stage. The Midian deposits in the central subzone are composed of massive light gray and pinkish detrital limestones, locally with cherty concretions and with fusulinid and tabulate remains. In the Kur-Teke and Kastanat-Dzhilga sections, rugose corals are represented by *Timorocarinophyllum zolkinae, Ufimia beirichi, Pavastephyllum (Tomasiphyllum) carcinophylloides, P. (Tomasiphyllum) carinatiseptatum, Iranophyllum (Laophyllum) pongouaense,* and *Waagenophyllum* (W.) *kueichowense.*[4] The thickness is 10–40 m. In the intermediate subzone, the Midian Stage corresponds to the Karasin Member composed of massive conglomeratelike limestones (0–25 m) with cherty inclusions and rugose corals, bryozoans, and brachiopods. The Rugosa assemblage found in the Igrim'yuz, Kutal', and Kara-Su sections consists of *Ufimia* aff. *persica, Pavastephyllum (Tomasiphyllum) carcinophylloides, Epiphanophyllum sinuosum, Iranophyllum (I.) splendens,* and *Lonsdaleiastraea* aff. *yamanbaensis.*

The Dzhulfian Stage. Deposits of the Dzhulfian Stage in the central subzone crown the section of massive limestones of the Upper Permian Kastanat-Dzhilga Formation, which yielded dendroid waagenophyllids *Waagenophyllum (W.) ramosum* and *Pseudohuangia pyzhjanovi.* These deposits (5–10 m) are virtually unseparable from the Midian strata forming the single sequence. In the intermediate subzone, the Dzhulfian Stage includes the upper layers of the Gana Formation (the Kutal' Member) and the lower Takhtabulak Subformation. The 20-m-thick Kutal' Member consists of thin-platy fine-grained and detrital limestones with cherty inclusions. These rocks are almost devoid of faunal remains; the only specimen of rugose coral *Waagenophyllum (W.) ramosum* was found in the Shindy section. In many sections of the intermediate subzone, conglomeratelike limestones of the Karasin Member overlying the Kutal' Member are missing, and uniform thin-platy limestones of the Deire and Kutal' members form the single sequence. The 10- to 20-m-thick lower Takhtabulak Subformation is composed of massive and thick-bedded organogenic, locally detrital, conglomeratelike and mainly algal limestones (10–25 m) with rugose and tabulate corals, brachiopods, and bryozoans. Corals are represented by *Waagenophyllum (W.) ramosum, W. (W.) kueichowense, W. (W.) virgalense, W. (Huayunophyllum) pamiricum, Amplexocarinia abichi,* and *Asserculina prima.*

The Dorashamian Stage. The deposits of the Dorashamian Stage (the middle Takhtabulak Subformation) are represented in the intermediate subzone (Kutal', Shur-Bulak, Kur-Istyk, Ak-Bura, Muzdu-Bulak, and Dunkel'dyk sections) by alternating tuffs, tuffites, tuffstones, and marly shales with limestone lenses and small bioherms (30–60 m) with diverse fauna of rugose corals, tabulates, brachiopods, bryozoans, and crinoids. The rugose assemblage includes *Ufimia* aff. *persimmetrica, Timorocarinophyllum zolkinae, Waagenophyllum (W.) ramosum, W. (W.) kueichowense, W. (W.)* aff. *virgalense, W. (W.) tachtabulasicum, W. (W.) crassiseptum, W. (Huayunophyllum) pamiricum, Pseudohuangia collucata, P. pyzhjanovi, Kabakovitchiella duplex,* and *K. arcuata.*

ANALYSIS OF THE ASSEMBLAGES

Rugose corals from Permian deposits of southeastern Pamirs are represented by 67 species (12 of them were determined in the open nomenclature) belonging to 24 genera and 12 families (Table 2). They occur in all the stages and in almost all the formations; they are, however, more common in the Kochusu, Gana, and Takhtabulak formations. Three stratigraphic units are defined: *Pseudoberbeekiellla–Timorocarinophyllum–Ufimia* (the Kochusu assemblage), *Iranophyllum–Pavastephyllum* (the Gana assemblage), and *Waagenophyllum (W.)–W. (Huayunophyllum)* (the Takhtabulak assemblage) zones, which are correlated with corresponding units in other regions.

The *Pseudoverbeekiella–Timorocarinophyllum–Ufimia* Zone corresponds to the Kochusu Formation of the Bolorian Stage. The zone contains the richest tetracoral assemblage (the Kochusu assemblage), which includes 28 species of 15 genera. The assemblage is mainly represented by solitary forms of the Plerophyllidae, Timorphyllidae, and Lophophyllidae families, among which endemic genera and species *Pseudoverbeekiellla variabilis, Timorocarinophyllum zolkinae,* and *Pentaphyllum permicum* are most abundant. Small colonial rugose corals *Calophyllum dobrolubovae* and *Pseudohuangia stoklini* also occur.

Most of these species are the same as the forms from Timor (Indonesia). Eight of them are reported from the Bituani Beds: *Pentaphyllum gerthi, Ufimia persimmetrica, U. variabilis,* and others such as *Verbeekiella gerthi* and *Wannerophyllum cristatum,* which occur in Timor not only in the Bituani Beds, but in the Basleo Beds as well, and also *Lophophyllidium spinosum* distributed in the Bituani, Basleo, and Amarassi beds. In addition to the last three species, the Kochusu assemblage includes seven species recorded in the Basleo Beds. These are *Ufimia kobayashi, Verbeekiella arboricalis,* and others. Hence, more than half of the species of the Kochusu assemblage are known from the Permian deposits of Timor. In other regions, comparable species are rare. Two species (*Cyathaxonia orientalis, Amplexocarinia socialis*) were recorded in the Artinskian Stage of the Urals, one species (*Ufimia alternata*) in the Chihsian Stage (the *Misellina* Zone) of China and in the Guadelupian Stage of Transcaucasia,

[4] The vesicular structure of septa allows this species to be referred to the subgenus *Tomasiphyllum* rather than to the *Sakamotosawanella* Minato et Kato, 1965.

and one species (*Cyathaxonia minima*) in the Darvazian Stage of Darvaz.

Deposits of the Shindy Formation of the Bolorian Stage yielded only small solitary tetracorals belonging to five species: *Amplexocarinia* aff. *subtilis, Wannerophyllum excedens, Kabakovitchiella* aff. *arcuata,* and *Euryphyllum cainodon* known from the Basleo Beds of Timor and *Yatsengia pachyteca* described from the Darvazian Stage of the northern Pamir.

Only two species cross the Lower–Upper Permian boundary in the southeastern Pamirs: *Timorocarinophyllum zolkinae* and *Wannerophyllum excedens.* The Kubergandy assemblage includes ten species of ten genera. One of these species is endemic (*Timorocarinophyllum zolkinae*), and three species were recorded in the Lower Permian: *Cyathaxonia multituberculata* in the Artinskian Stage of the Urals, *Pseudohuania* aff. *tabellata* in the *Pseudofusuliina* Zone of Iran, and *Yatsengia asiatica* in the *Pseudofusuliina* Zone of Iran and *Misellina* Zone of South China. Two species are known from both the Lower and Upper Permian: one is *Amplexocarinia heimoi* described from the Artinskian deposits of the Urals, Darvazian Stage of the northern Pamirs, Lower Permian deposits of the Carnian Alps, Basleo Beds of Timor, and *Parafusulina stricta* Zone of southern Primor'e; another form *Praewentzelella magnifica* is reported from the *Pseudofusulina–Parafusulina* zones of Iran. Beyond the limits of the southeastern Pamirs, four species occur only in the Upper Permian: *Wannerophyllum excedens* and *Kabakovitchiella duplex* are recorded in the Basleo Beds of Timor, *Pavastephyllum (Tomasiphillum) simplex* occurs in the *Parafusulina–Neoschwagerina* Zone of Iran and Eastern Caracorum, and *Duplophyllum tenuiseptatum* is reported from the Basleo Beds of Timor and *Parafusulina* Zone of southern Primor'e.

Thus, the Late Permian forms distinctly prevail in the Kubergandian Stage.

The *Pavastephyllum–Iranophyllum* Zone (the Gana assemblage) unites the Murgabian and Midian stages. The tetracoral assemblage consists of 17 species of 8 genera, and 50% of its forms are represented by waagenophyllids. Most characteristic of them are solitary forms of *Pavastephyllum* and *Iranophyllum* genera: *I. splendens* and *P. carcinophylloides* from the *Pseudofusulina* Zone of Iran, *I. amigdalophylloideum* from the *Parafusulina* Zone of South China, and species *I. pongouaense* and *P. carcinatoseptatum* from the *Neoschwagerina* Zone of Laos and Cambodia. Massive waagenophyllids include *Lonsdaleiastraea typica* occurring in the *Parafusulina–Neoschwagerina* zones of Eastern Caracorum and relative species *L. yamanbaensis* recorded in the Maidzhuru Beds and *Yabeina* Zone of Japan. Dendroid waagenophyllids are represented by *Waagenophyllum kueichowense*, which is known in the above-mentioned zones of Iran, Iraq, and China, as well as in the *Metadoliolina lepida* Zone of southern Primor'e. In South China, this form occurs up to the Chansingian Stage. Four *Ufimia* species are common in Timor (Indonesia), and they indicate the faunal exchange between these regions; three other species are endemic: *Timorocarinophyllum zolkinae, Iranophyllum igrimjusicum,* and *Epiphanophyllum sinuosum.*

The *Waagenophyllum (W.)– W. (Huayunophyllum)* Zone (the Takhtabulak assemblage) is correlative with the Dzhulfian and Dorashamian stages.

The zonal assemblage consists of 15 species of 7 genera; seven species out of this number are endemic. Most of the species (8) are represented by colonial waagenophyllids, *Waagenophyllum (W.) kueichowense* and *W. (W.) virgalense* included. The latter is widespread in the *Neoschwagerina–Yabeina* zones of the Tethyan realm (Yugoslavia, Mongolia, Burma, Pakistan, Afghanistan, Indochina) and Japan, as well as in the *Metadoliolina lepida* Zone of southern Primor'e. Other waagenophyllids are represented by six endemic species: *Pseudohuangia collucata, P. pyzhjanovi, W. (W.) tachtabulasicum, W. (W.) crassiseptum, W. (W.) ramosum,* and *W. (Huayunophyllum) pamiricum.* The presence of the *Huayunophyllum* genus suggests faunal exchange with the Himalayas and South China basins (Xu, 1984; Sheng, *et al.,* 1984), where it is characteristic of the Chihsian deposits. Species of the *Amplexocarinia, Asserculinia,* and *Kabakovitchiella* genera are characteristic of the Basleo deposits of Timor (Indonesia).

Thus, the faunal exchange between the southeastern Pamirs and Indonesian basin (Timor Island) persisted during the entire Permian and ceased in the Late Permian, while the connection with the eastern part of the Tethys became simultaneously wider. The increase in the share of endemic species in the Dorashamian time suggests isolation of the basin.

TAXONOMIC INTERPRETATION. MORPHOLOGY

Certainly, the most interesting taxonomic and morphologic aspects of rugose corals from the southeastern Pamirs are considered below. Some families can be united if we take into consideration the trabeculae location in the septal plane. On the basis of the fan-shaped trabeculae arrangement, the Verbeekiellidae family is considered as a junior synonym of the Timorphyllidae family, whereas the Wannerophyllidae family is considered to be a junior synonym of the Lophophyllidae (=Lophophyllididae) family based on the monoclinal arrangement of trabeculae in the septal plane (Fig. 3). Hill (1981) considers genera *Verbeekiella* and *Wannerophyllum* (Plate 4, 9) to be synonymous; however, the different microstructure of septa allows them to be classified not only as independent genera but also as representatives of different families.

The difference between similar *Pseudoverbeekiella* and *Verbeekiella* genera is in the counterseptum of the former genus, which evidently develops simulta-

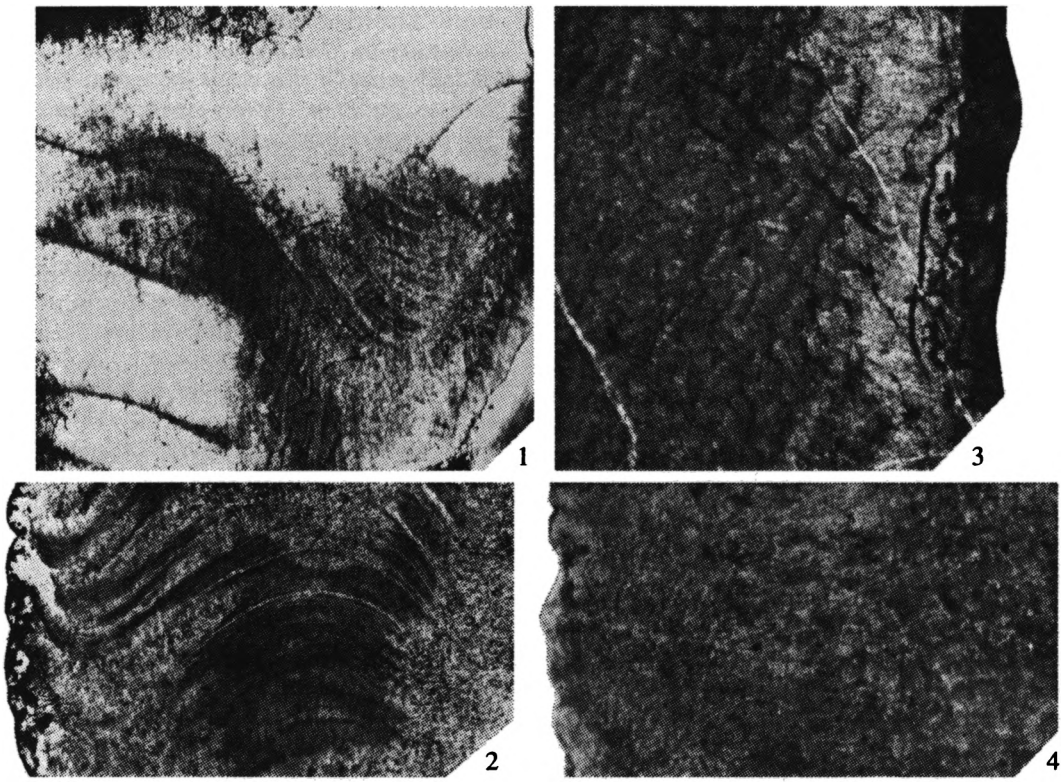


Fig. 3. Longitudinal cross sections of septa: (1) *Timorocarinophyllum zolkinae* Iljina, 1997. Holotype 2376/883 PIN, fan-shaped arrangement of trabeculae in the septal plane, thin section 1, the Kur-Istyk River, left slope of the second upper left tributary, 1.5 km from the mouth, the Lower Permian, Bolorian Stage, upper Kochusu Subformation, magn. 20; (2) *Pseudoverbeekiella variabilis* Iljina, 1997, specimen 2376/30 PIN, fan-shaped arrangement of trabeculae in the septal plane, thin section 2, the Igrim'yuz River, left side near the mouth, the Lower Permian, Bolorian Stage, lower Kochusu Subformation, magn. 20; (3) *Wannerophyllum cristatum* (Gerth, 1921), specimen 2376/572 PIN, monoclinal arrangement of trabeculae in the septal plane, thin section 6, the Kur-Teke River upper reaches, second left tributary, the Lower Permian, Bolorian Stage, lower Kochusu Subformation, magn. 50; (4) *Lophophyllidium* (*Lophbillidium*) *martini* Schouppe et Stacul, 1955, specimen 2376/218 PIN, monoclinal arrangement of trabeculae in the septal plane, thin section 7, the Mamazair-Bulak locality, 1 km south of the 467-km mark of the Pamirian road, the Lower Permian, Bolorian Stage, lower Kochusu Subformation, magn. 50.

neously with the main one, because it is elongated already at the early neanic stage and directly joins the axial columella (becoming shorter only in the calyx). Due to this feature, the median plate appears at the initial stages of columella growth, and then the honeycomb–tubular structure develops. Revision of the *Verbeekiella* ontogenesis (Fedorovski, 1986) showed that the underdeveloped counterseptum in the tetracorals of this genus appearing later than other protosepta is equal in length to metasepta and does not take part in the formation of the axial structure during the entire ontogenesis. Axial ends of only four protosepta, two lateral and two neighboring the counterseptum, join in the center and give rise to the formation of the axial structure of corals of this genus. The *Timorocarinophyllum* genus is defined based on the carinae development.

I was able to point out for the first time that the septum microstructure of *Lophophyllidium* (*Lophbillidium*) *spinosum* (Plate 4, 6a; Fig. 4, 1) consists of a single row of small simple trabeculae, whereas *L.* (*Lophbillidium*) *martini* (Plate 4, 8; Fig. 4, 2) has a septum consisting of complicated trabeculae; in the latter, only septa of the early stages are represented by a single row of simple trabeculae. The complicated septum microstructure (combination of simple and complex trabeculae) is noted in the Dorashamian species *Waagenophyllum crassiseptum* and *Pseudohuangia collucata* (Plate 3, 1c, 7b). Analysis of the tetracoral microstructure leads to the conclusion that the trabeculae number and, to a certain extent, their dimensions, inclination angle in monoclinal series, and the number of rows are features of the species rank, whereas the arrangement of trabeculae in the septum plane (fan-shaped or monoclinal) and the trabecula type (holocanth, rhabdocanth, monocanth, simple, and complex) are characteristics of the genus and family rank.

DESCRIPTION OF RUGOSE CORALS

Subclass Rugosa

Order Stauriida

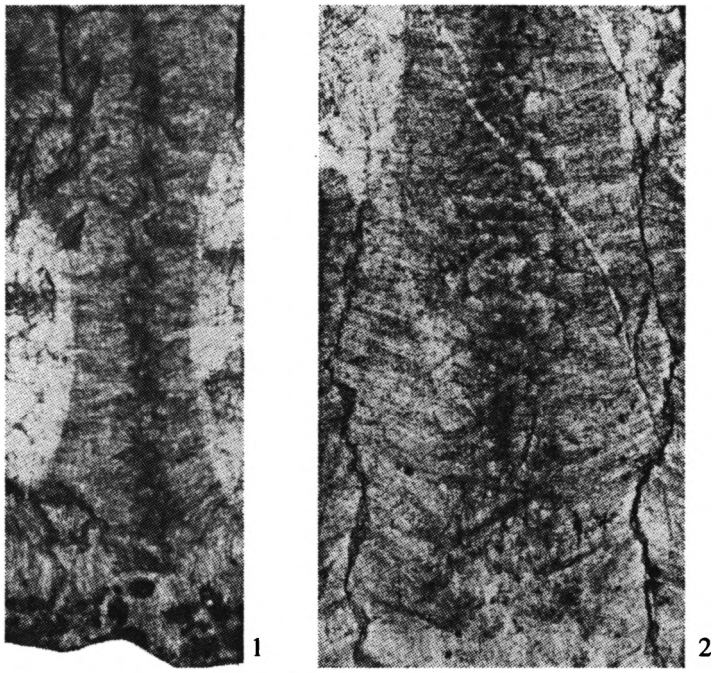


Fig. 4. Transverse cross sections of septa: (1) *Lophophyllidium* (*Lophbillidium*) *spinosum* Schouppe et Stacul, specimen 2376/525 PIN, transverse cross section of septum, thin section 2, the Mamazair-Bulak locality, 1 km south of the 467-km mark of the Pamirian road, the Lower Permian, Bolorian Stage, lower Kochusu Subformation, magn. 50; (2) *Lophophyllidium* (*Lophbillidium*) *martini* Schouppe et Stacul, 1955, specimen 2376/218 PIN, transverse cross section of septum, thin section, the locality and stratigraphic position, magn. 50.

Family Pleurophyllidae Koker, 1924

Genus *Pentaphyllum* Konick, 1872

Pentaphyllum permicum Iljina, 1997

Plate 1, 1–3. *Pentaphyllum permicum*: Ilyina, 1997, p. 134, Plate 1, 1

Etymology. The name of the species is derived from the term "Permian System."

Holotype. PIN RAS, specimen 2376/367; southeastern Pamirs, southern part of the Kurteke River valley, upper left tributary; Lower Permian, Bolorian Stage, lower Kochusu Subformation.

Description. Small cone-shaped corals with distinct rugae on the outer surface (Plate 1, 1b). When the diameter is 11.5 mm, the septum number equals (25–29) × 2 (Plate 1, 1a, 3b). Five protosepta (main, lateral, and neighboring the counterseptum) are most elongated and clavate-thickened, while the counterseptum is less elongated; its length varies from 2/3 to 5/6 of the length of joining protosepta. Median protosepta are elongated. Interseptal spaces are small and located only along the periphery of tabularium. Septa of the second order are commonly developed within the wall only at the fibrous stage (Plate 1, 1b), but in the upper part of the calyx with the developed trabecular plates, they protrude in the form of small crests (Plate 1, 3b). The number of septa in main quadrants is 3–4 less than in counter quadrants. The bases of trabecular laminae consisting of small (0.07 mm) simple trabeculae are separated from the wall and appear to be submerged into the fibrous mass as deep as 1/2–2/3 of the wall thickness or only joined to it. The inclination angle of trabeculae in the septum plane is 68° (Plate 1, 1d). The wall is up to 2 mm thick. The zigzag location of fibers is determined by the wavy surface of layers of the fiber growth (Plate 1, 1b). Tabulae are thin and convex (Plate 1, 1c). During the ontogenesis, the counterseptum becomes shorter, and interseptal spaces appear (Plate 1, 2, 3a).

Comparison. The species differs from *P. australe* (Hinde, 1890) in the clavate form of five elongated protosepta, in smaller main quadrants, and in the later development of trabeculae (in *P. australe*, they reach epitheca). It also differs from *P. radiciforme* (Gerth, 1921) by the greater thickness of five protosepta.

Distribution. Lower Permian, Bolorian Stage, lower Kochusu Subformation of the southeastern Pamirs.

Material. One specimen from the Igrim'yuz locality, 16 specimens from the Kur-Teke locality, and 12 specimens from the Mamazair-Bulak locality.

Family Timorphyllidae Soskina, 1941

Genus *Timorocarinophyllum* Ilyina, 1997, p. 134

Etymology. The name of the genus is derived from "*Timorphyllum*" and Latin "carina" (keel).

Type species: *T. zolkinae* Illjina.

Diagnosis. Solitary corals without dissepiments. Septa are of two orders. The counterseptum is elon-

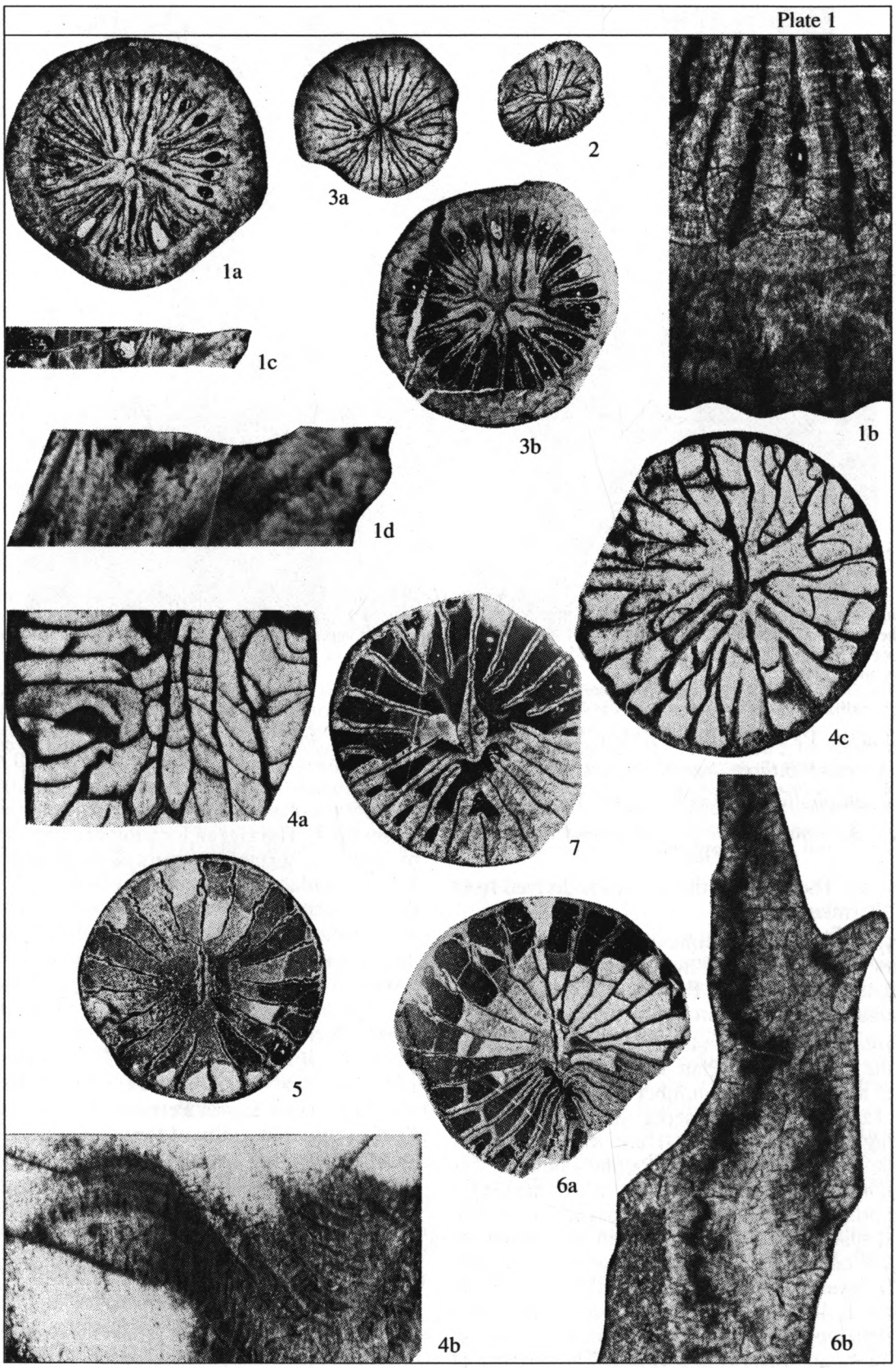


Plate 1
1a
1c
1d
3a
2
3b
1b
4c
4a
7
5
6a
4b
6b

gated; one or several pinched ends of septa of the first order join its axial end. Sporadic carinae are developed on the lateral surfaces of septa. Trabeculae are arranged in the septum plane in the fan-shaped mode.

Species composition. The genus is monospecific.

Comparison. The genus differs from *Wtimorphyllum* in that it has carinae.

Distribution. The Lower Permian, Bolorian Stage; Upper Permian,

Kubergandian–Dorashamian stages.

Timorocarinophyllum zolkinae Iljina, 1997

Plate 1, 4–7; Fig. 3, 1

Timorocarinophyllum zolkinae Ilyina, 1997, p. 134, Plate 1, 2; Fig. 4, 1

Etymology. The species is named in honor of A.I. Zolkina, a Russian paleontologist.

Holotype. PIN PAS, specimen 2376/883; southeastern Pamir, the second (upstream) tributary of the Kur-Istyk River, 1.5 km above the mouth; Lower Permian, Bolorian Stage, upper Kochusu Subformation.

Description. Corals are narrow-conic, subcylindric in the upper part, and 5–6 cm high. The calyx is deep. In specimens with a diameter of 10–20 mm, the number of septa is (20–25) × 2. Metasepta reach or almost reach the pseudocolumella, being arranged in the radial mode in the counterquadrants and, sometimes, in the feather-like mode along the main septum in the main quadrants. The counterseptum exceeds in length the coral radius and is thickened at the axial end and conjoined with one to three pinched axial ends of the septum (Plate 1, 4c, 5, 6a, 6b, 7). The main septum is shortened and located in the fossula. The lateral septum surfaces of the first order bear carinae often conjoined with the septum through stereoplasma (Plate 1, 5, 6a–b, 7). Septa of the second order are developed within the epitheca only at the fibrous stage; locally, they correspond in form to semicircular projections (Plate 1, 4c). Septa of the first order consist of small simple trabeculae (0.05 mm) arranged in the fan-shaped mode in the septum plane (Plate 1, 4b; Fig. 3, 1). The wall is thin, averaging 0.3 mm. Tabulae are gently convex, thin, splitted along the periphery, and elevated in the columella area (Plate 1, 4a). Septa of main quadrants are often strongly thickened by stereoplasma and join each other (Plate 1, 7), often up to the calyx base.

Distribution. The Lower Permian, Bolorian Stage, Kochusu Formation; the Upper Permian, Kubergandian Stage, Kubergandy Formation; Murgabian Stage, Dzhamantal Member; Doroshamian Stage, middle Takhtabulak Subformation of the southeastern Pamir.

Material. The Bolorian Stage: 4 specimens from the Igrim'yuz locality, 15 specimens from the Kur-Istyk locality, 5 specimens from the Mamazair-Bulak locality, 1 specimen from the Chakobai locality, 2 specimens from the Southern Ak-Arkhar locality, 2 specimens from the Sulyu-Dzhilga locality, and 3 specimens from the Ak-Bura locality. The Kubergandian Stage: 3 specimens from the Kur-Teke locality and 3 specimens from the Kutal' locality. The Murgabian Stage: 2 specimens from the Kutal' locality and 2 specimens from the Kur-Teke locality. The Dorashamian Stage: 4 specimens from the Kur-Istyk locality.

Genus *Pseudoverbeekiella* Iljina, 1997

Pseudoverbeekiella Ilyina, 1997, p. 135

Etymology. The name of the genus is derived from the Greek "pseudo" (seeming) and the *Verbeekiellla* genus name.

Type species: *P. variabilis* Iljina, 1997.

Diagnosis. Solitary corals without dissepimentarium. Septa are of two orders. The counterseptum is elongated with the honeycomb-tubular axial columella at its axial end. The main septum is shortened. Septa consist of small simple trabeculae arranged in the septa planes in the fan-shaped mode. Tabulae are convex.

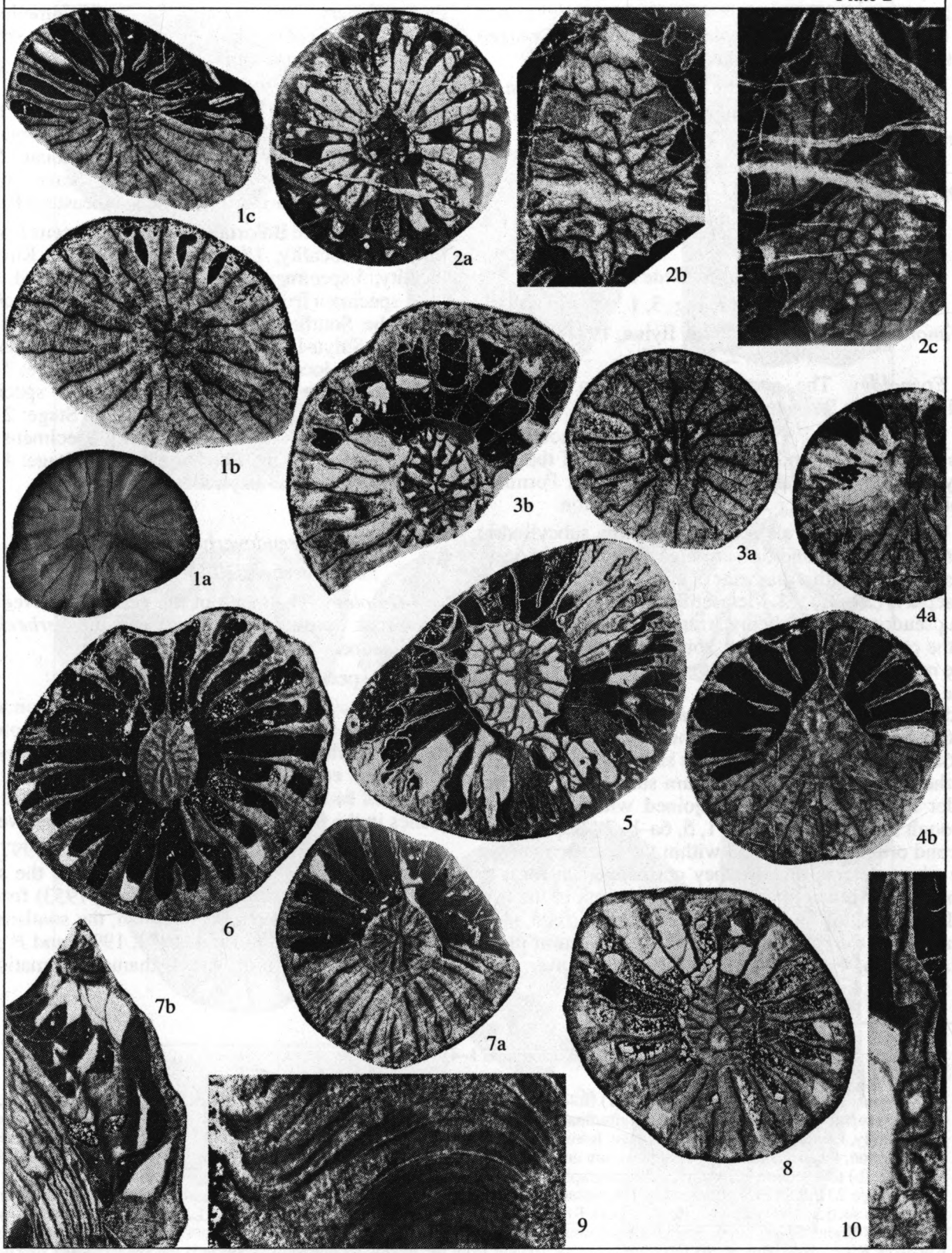
Species composition: P. variabilis Iljina, 1997 from the Bolorian Stage of the Lower Permian, the southeastern Pamir; *P. ussurica* (Fomitchev, 1953) from the Midian Stage of the Upper Permian, the southern Primor'e; *P. australis* (Fluegel, 1972, 1990) and *P. parva* (Fluegel, 1972) from the Dzhamal Formation of

Transverse and longitudinal cross sections of corallites (Plates 1–4)

Plate 1.

(1–3) *Pentaphyllum permicum* Iljina, 1997. (1) Holotype 2376/367 PIN, thin section 7, transverse cross sections of corallite (1a) ×4, walls and septa (1b) ×20; thin section 5, longitudinal cross section of corallite (1c) ×4, and septum (1d) ×15, the Kurteke River, upper left tributary, Lower Permian, Bolorian Stage, lower Kochusu Subformation. (2) specimen 2376/573 PIN, thin section 2, transverse cross section, locality and stratigraphic position are the same, ×6. (3) specimen 2376/258 PIN, transverse cross sections, (a) thin section 5, (b) thin section 1, locality and stratigraphic position are the same, ×4. (4–7) *Timorocarinophyllum zolkinae* Iljina, 1997. (4) Holotype 2376/883 PIN, thin section 1, longitudinal cross sections of corallite (4a) ×4, septum and carina (4b), ×20 (4c) transverse cross section, thin section 2, the Kur-Istyk River, left wall of the second (upper) left tributary, 1.5 km from the mouth, Lower Permian, Bolorian Stage, upper Kochusu Subformation. (5) specimen 2376/286 PIN, transverse cross section, thin section 15, the Kur-Istyk River, right bank of the third left tributary, 0.6 km from the mouth, stratigraphic position is the same, ×4. (6) specimen 2376/324 PIN, thin section 7, transverse cross sections of corallite (6a) ×3, and septum with carinae (6b) ×50, the Kur-Istyk River, right bank of the third left tributary, 0.25 km upstream of the mouth, Lower Permian, Dorashamian Stage, middle Takhtabulak Subformation. (7) specimen 2376/831 PIN, transverse cross section, the Mamazair River, 1 km south of the 467-km mark of the Pamirian road, Lower Permian, Bolorian Stage, lower Kochusu Subformation, ×4.

Plate 2



eastern Iran and the Shaksgam Formation of northern Caracorum (China).

Comparison. The genus in question differs from the similar genus *Verbeekiella* Penecke, 1908, revised by Fedorowski (1986), in having an elongated counterseptum with the axial structure developed at the axial end.

Distribution. The Lower Permian of West Australia, eastern Iran, northern Caracorum (China), and southeastern Pamir; the Upper Permian of the southern Primor'e region.

Discussion. The comparison of the Pamirian corals of the *Pseudoverbeekiella* genus with *Verbeekiella australis* and *V. gerthi* (non *V. gerthi* Schouppe et Stacul, 1955) described by Fluegel (1972, 1990) suggests that we are dealing here not only with representatives of a single genus but with very close species. All these forms have the elongated counterseptum with the honeycomb-tubular axial structure at its end. The difference between illustrated specimens is due to different stages of development of these structures, as was shown by the thorough study of the Pamirian corals.

Inasmuch as Fluegel did not give a detailed description of the morphology, ontogeny, microstructure, and variability of the studied species, an integration of these and Pamirian forms into a single species would be premature, though they certainly belong to a single group. As for *P. australis* (Fluegel) and *P. parva* (Fluegel), in my opinion, they are synonymous.

The Pamirian forms slightly differ because of greater height and diameter, as if their development were slower in comparison with the corals described by Fluegel. For instance, *P. parva* specimens with a diameter of 12–13 mm have 26 septa and the developed honeycomb axial structure, while the Pamirian forms bear these features only when their diameter is as large as 24 mm.

Pseudoverbeekiella variabilis Iljina, 1997

Plate 2, 1–10; Fig. 3, 2

Pseudoverbeekiella variabilis: Ilyina, 1997, p. 135, Plate 1, 3–7; Fig. 4, 2

Etymology. The name of the species is derived from Latin "variabilis" (changeable).

Holotype. PIN RAS, specimen 2676/225, the southeastern Pamir, Mamazair-Bulak area; the Lower Permian, Bolorian Stage, lower Kochusu Subformation.

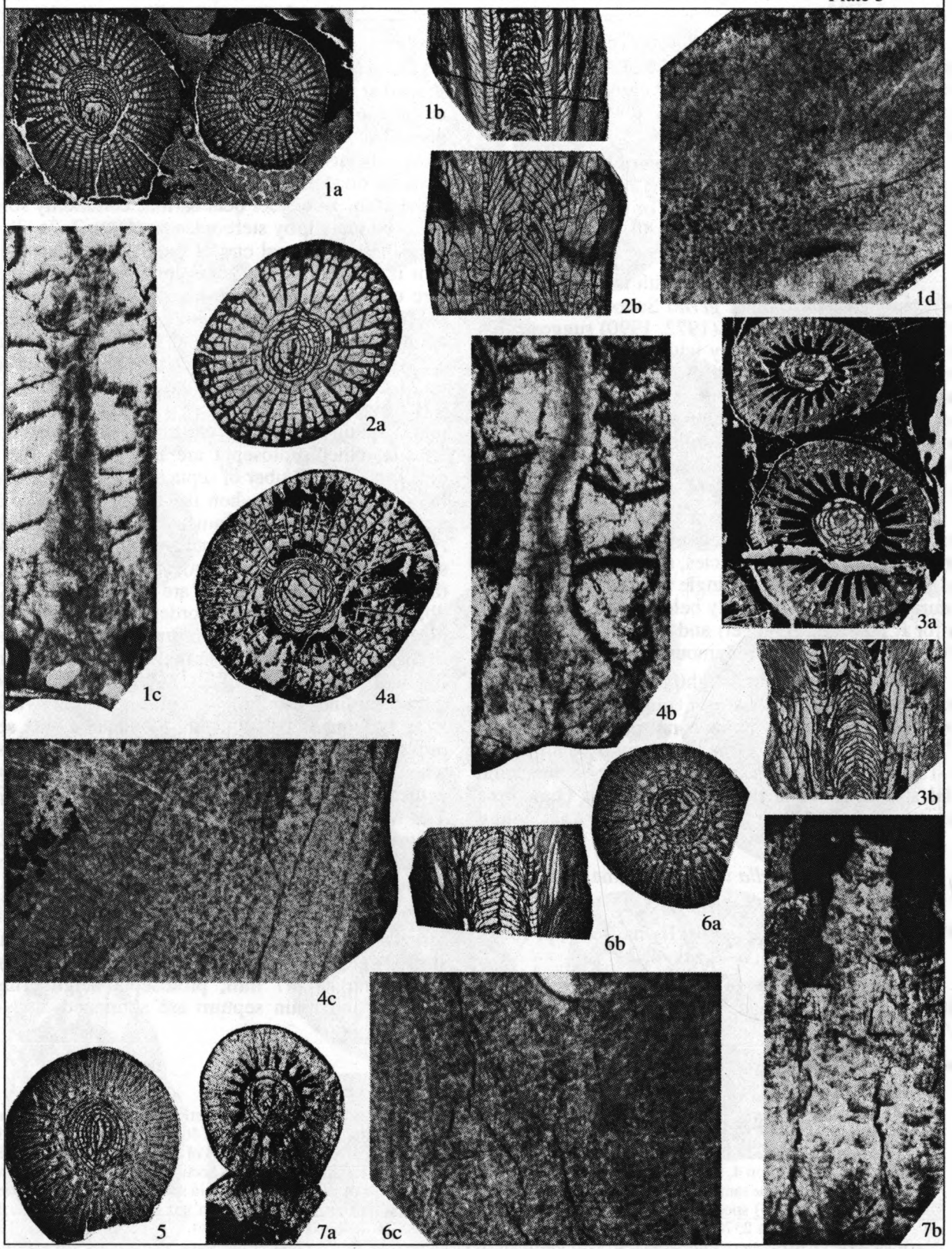
Description. These narrow-conic corals are horn-shaped at the early stage and commonly subcylindrical at the adult stage. The height is 5–9 cm. The calyx is deep. The number of septa is 20 × 2 in specimens with the diameter of 10–12 mm and 25 × 2 in specimens with the diameter of 23–24 mm. The lateral surfaces of septa of the first order bear carinae commonly soldered with the septum by stereoplasma (Plate 2, 2a, 3b, 5, 6, 7a, 8, 10). The axial end of the elongated counterseptum is marked by the honeycomb-tubular axial structure of oval, droplike, or occasionally rounded form. Its dimensions are quite variable. For instance, if the coral diameter is 12 mm, the axial structure may be 3.3 × 4 or 3.9 × 4 mm in size; when the diameter makes up 24–25 mm, dimensions of the axial structure vary from 4 × 7 mm (Plate 2, 2c) to 6.6 × 7 mm (Plate 2, 5). The main septum is shortened and commonly situated in the fossula; other protosepta are indistinct. In the main quadrants, the number of septa is two or three less than in counterquadrants when the diameter is 10–12 mm and four less when the diameter increases up to 24 mm. At the initial stages, all septa are strongly thickened by stereoplasma (Plate 2, 1a, 4a); in adult specimens, only the septa of main quadrants are thickened (Plate 2, 1c, 4b, 7a, 8). Septa of the first order are composed of simple trabeculae (Plate 2, 10) arranged in the septa planes in the fan-shaped mode but inclined to the center near the wall (Plate 2, 9; Fig. 3, 2). At the initial stage, the trabecular line is straight, whereas it is often twisting in adult specimens (Plate 2, 3b, 8). Septa of the second order are developed inside the wall at the fibrous stage; when trabecular laminae are developed, they protrude in the form of small carinae only in the calyx (Plate 2, 3b). The wall is thin. Tabulae rise in the zone of the axial columella to form dome-shaped elements (Plate 2, 7b). Sporadic dissepiments are sometimes developed on the convex side of the lower curved part.

Ontogenesis. Initial ontogenetic stages are not preserved; however, it is seen that the counterseptum is strongly elongated even in more mature specimens with a diameter of 4.7 mm; protosepta neighboring the counter- and main septum are shortened, and every

Plate 2.

(1–10) *Pseudoverbeekiellla variabilis* Iljina, 1997. (1) specimen 2376/826 PIN, transverse cross sections (1a) polished section 3, ×7; (1b) thin section 4, ×7; (1c) thin section 5, ×4, the Mamazair-Bulak locality, 1 km south of the 467-km mark of the Pamirian road. (2) holotype 2376/225 PIN, (2a) transverse cross section, thin section 2, ×3; transverse sections of the axial structure: (2b) thin section 3, (2c) thin section 4, ×10, the same locality. (3) specimen 2376/406 PIN, transverse cross sections: (3a) thin section 3, ×3; (3b) thin section 9, ×4, the same locality. (4) specimen 2376/219 PIN, transverse cross sections: (4a) thin section 1, (4b) thin section 2, the same locality, ×5. (5) specimen 2376/15 PIN, transverse cross section, thin section 1, the Igrim'yuz River, left bank near the mouth, ×3. (6) specimen 2376/20 PIN, transverse cross section, thin section 1, the same locality, ×4. (7) specimen 2376/33 PIN, (7a) transverse cross section, thin section 4; (7b) longitudinal cross section, thin section 5, the same locality, ×4. (8) specimen 2376/34 PIN, transverse cross section, thin section 1, the same locality, ×4. (9) specimen 2376/30 PIN, longitudinal cross section of septum, thin section 2, the same locality, ×20. (10) specimen 2376/249 PIN, transverse cross section of septum with carinae, thin section 1, the Mamazair-Bulak locality, 1 km south of 467-km mark of the Pamirian road, ×10. All specimens are from the Lower Permian, Bolorian Stage, lower Kochusu Subformation.

Plate 3



quadrant contains two metasepta. All the septa are strongly thickened by stereoplasma and join each other (Plate 2, 1a). When the diameter of coral increases to 6 mm, the axial columella begins to develop at the axial end of the elongated counterseptum (Plate, 3a). Later, the columella develops owing to the joining of pinched axial ends of septa and the appearance of supplementary laminae (Plate 2, 1b). In specimens with a diameter of 7–10 mm, the median lamina remains visible a little while (Plate 2, 1b, 2a); later, it disappears (Plate 2, 2b), and a honecomb-tubular structure gradually forms in the center of the axial columella, until the diameter increases to 24 mm (Plate 2, 2b, 5). Sometimes, the counterseptum remains connected with the axial columella even in the calyx (Plate 2, 2b, 4b, 6, 7a, 8); commonly, it becomes separated and almost equal to other septa in length.

Comparison. The new species differs from relative forms *P. parva* (Fluegel, 1972) and *P. australis* (Fluegel, 1972) in having larger dimensions, in a later development of the honeycomb-tubular structure of the axial columella, and in the septa number reaching the maximum only in specimens with a diameter of 23–24 mm.

Distribution. The Lower Permian, Bolorian Stage, Kochusu Formation of the southeastern Pamir.

Material: 19 specimens from the Igrim'yuz locality, 21 specimens from the Mamazair locality, 1 specimen from the Kur-Istyk locality, and 1 specimen from the upper Chakobai Subformation.

Family Waagenophyllidae Wang, 1950

Subfamily Waagenophyllinae Wang, 1950

Genus *Pseudohuangia* Minato et Kato, 1965

Pseudohuangia collucata Iljina, 1997
Plate 3, 1

Pseudohuangia collucata: Ilyina, 1997, p. 135, Plate 1, 8; Fig. 5, 3

Etymology. The name of the species is derived from Latin "*collucata*" (delicate).

Holotype. PIN RAS, specimen 2376/926; the southeastern Pamir, Shur-Bulak-Sai, left-hand side near the mouth; Upper Permian, Dorashamian Stage, middle Takhtabulak Subformation.

Description. Large cylindrical corallites 9–11 mm across. Septa are thick, (28–31) × 2 (Plate 3, 1a). Septa of the first order are long and commonly straight; they do not reach the axial collumella and are most elongated in the peripheral and axial parts (Plate 3, 1b). Septa of the second order are 3/4 shorter than septa of the first order. Septal microstructure is trabecular; large (0.075–0.1 mm) simple trabeculae are arranged in the median part of septa in the two-row mode. In the peripheral and axial parts (Plate 3, 1b, 1d), septa of the first order are composed of complex trabeculae (0.8–0.85 mm). The outer wall is thin (0.1-0.07 mm). The dissepimentarium consists of two to three rows of small rounded dissepiments in the outer zone and of two rows of large elongated and steeply inclined dissepiments in the inner zone (Plate 3, 1b). Tabulae are gently inclined and split near the axial collumella, where they are subhorizontal and even elevated. The axial columella is large, rounded, delicate, and occupies one third of the corallite diameter. It consists of densely arranged (5–6 in 1 mm) slightly convex dome-shaped axial tabulae combined with the thin median lamina that crosses the entire columella and with numerous discrete radial laminae.

Comparison. The described species differs from *P. tabellata* (Douglas, 1936) in larger dimensions and thicker septa with the clavate-thickened axial ends (in *P. tabellata*, the latter are sharpened). It also differs from *P. pyzhjanovi* Iljina in more elongated septa of the second order, denser arrangement of tabulae, and a greater number of septa.

Plate 3.

(1) *Pseudohuangia collucata* Iljina, 1997.

Holotype 2376/926 PIN, (1a) transverse cross section, thin section 1, ×3; (1c) transverse cross section of septum, thin section 1, ×20; thin section 2, longitudinal cross sections of corallite (1b) ×3, and septum (1d) ×50, the Shur-Bulak River, left side near the mouth, Upper Permian, Dorashamian Stage, middle Takhtabulak Subformation.

(2) *Pseudohuangia pyzhjanovi* Iljina, 1997.

Holotype 2376/334a PIN, (2a) transverse cross section, thin section 2; (2b) longitudinal cross sections, thin section 2, the Kutal' River middle reaches, 1 km south of the height 3895.3 m, Upper Permian, Dorashamian Stage, middle Takhtabulak Subformation, ×4.

(3, 4) *Waagenophyllum* (*Waagenophyllum*) *tachtabulasicum* Iljina, sp. nov.

(3) holotype 2376/344 PIN, (3a) transverse cross section, thin section 1; (3b) longitudinal cross sections, thin section 2, the Kur-Istyk River, fourth left tributary, Upper Permian, Dzhulfian Stage, lower Takhtabulak Subformation. (4) specimen 2376/955a PIN, transverse cross sections of corallite (4a) ×4, and septum (4b) ×50; (4c) longitudinal cross sections of septum, thin section 3, ×50, the Muzdu-Bulak River upper reaches, left bank, Upper Permian, Dorashamian Stage, middle Takhtabulak Subformation.

(5–7) *Waagenophyllum* (*Waagenophyllum*) *crassiseptum* Iljina, 1997.

(5) Holotype 2376/770 PIN, transverse cross section, thin section 1, the Kutal' River, Zou-Tash locality, height 389.3, Upper Permian, Dorashamian Stage, middle Takhtabulak Subformation, ×5. (6) specimen 2376/673 PIN, (6a) transverse cross section, thin section 2, ×5; thin section 3, longitudinal cross sections of corallite (6b) ×5, and septum (6c) ×50, locality and stratigraphic position are the same. (7) specimen 2376/193 PIN, thin section 2, transverse cross sections of corallite (7a) ×5, and septum (7b), the Ak-Bura River, left side, 6 km from the mouth, Upper Permian, Dorashamian Stage, middle Takhtabulak Subformation.

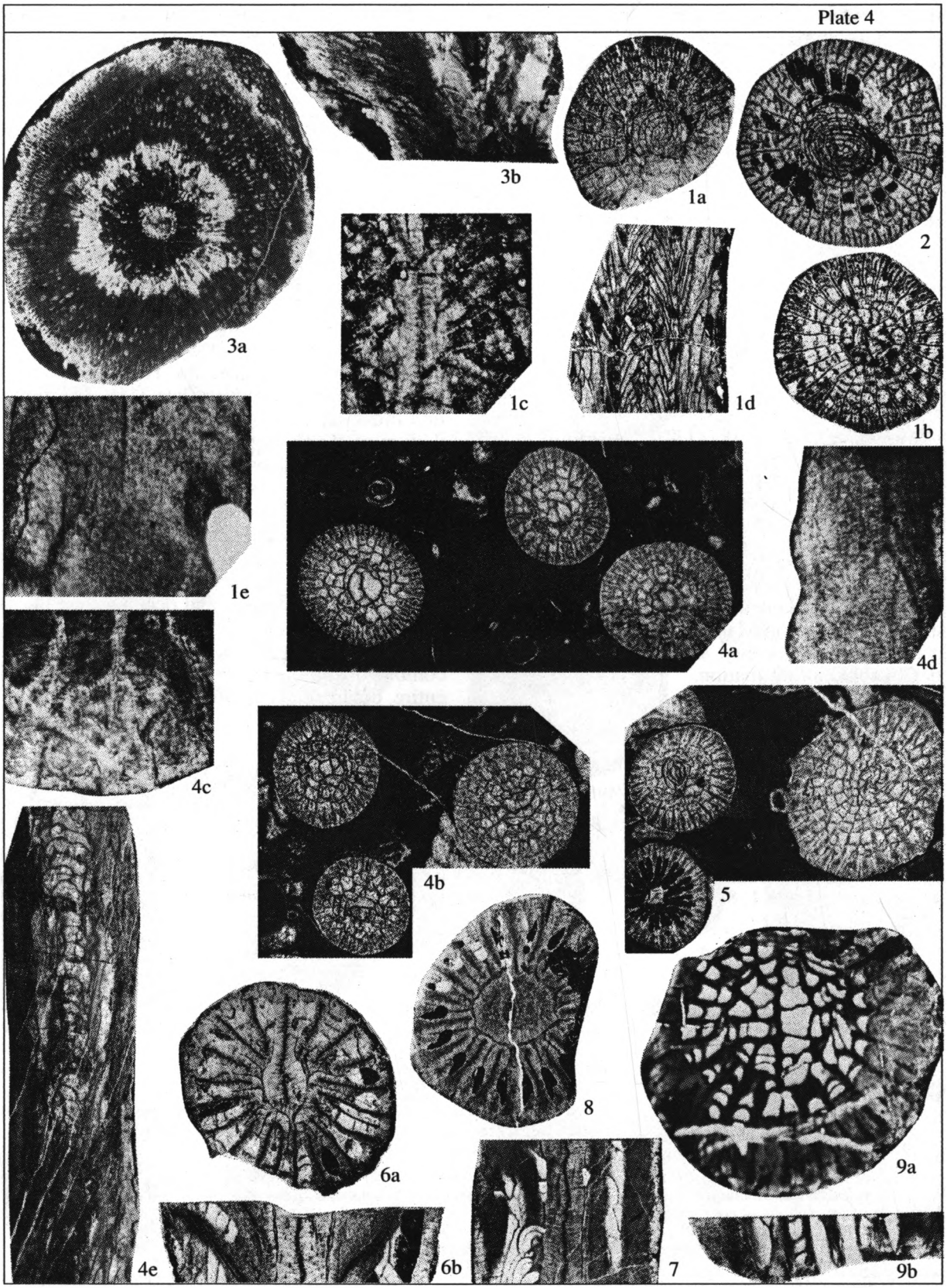


Plate 4
3b
1a
2
3a
1c
1d
1b
1e
4a
4d
4c
4b
5
4e
8
6a
9a
6b
7
9b

Distribution. The Upper Permian, Dorashamian Stage, middle Takhtabulak Subformation of the southeastern Pamir.

Material. Three specimens from the Kutal' locality and four specimens from the Shur-Bulak locality.

Pseudohuangia pyzhjanovi Iljina, 1997
Plate 3, 2

Pseudohuangia pyzhjanovi Ilyina, 1997, p. 135, Plate 1, 9

Etymology. The species is named in honor of I.V. Pyzh'yanov, a Russian specialist in corals.

Holotype. PIN RAS, specimen 2376/334a; the southeastern Pamirs, Kutal' River middle reaches, 1 km to the north of the height 3895.3 m; the Upper Permian, Dorashamian Stage, middle Takhtabulak Subformation.

Description. Corallites are subcylindrical, 8–10 mm in diameter. Septa of two orders are thin and (24–25) × 2 in number. Septa of the first order are wedge-shaped; commonly, they do not reach the axial columella. Septa of the second order are twice as short as those of the first order (Plate 3, 2a). Septa are built of small simple trabeculae (0.05 mm) inclined in the septum plane at the angle of 70°. The dissepimentarium is narrow and composed of two to three rows of small rounded and, locally, elongated dissepiments. The wide tabularium occupies about half of the corallite radius; tabulae are represented by strongly inclined clinotabulae located at a distance of 0.5–1 mm from each other. The axial columella has an oval form and occupies one third of the corallite diameter; it consists of the median laminae, which crosses the entire columella, and numerous radial laminae. Axial tabulae, three to four per mm, are pyramidal (Plate 3, 2b). At the boundary between dissepiment zones and tabulae, there is an inner wall insignificantly thickened by stereoplasma. The outer wall is thin (0.15 mm) and consists of epitheca and contiguous peripheral septal ends.

Comparison. The species differs from *P. tabellata* (Douglas, 1936) in the sharpened form of axial tabulae and well-developed steeply inclined but rare clinotabulae. Comparison with *P. collucata* is given above.

Distribution. The Upper Permian: Dzhulfian Stage, lower Takhtabulak Subformation; and Dorashamian Stage, middle Takhtabulak Subformation of the southeastern Pamir.

Material. Dzhulfian Stage: one specimen from the Kur-Teke locality; Dorashamian Stage: three specimens from the Kutal' locality and four species from the Kur-Istyk locality.

Genus *Waagenophyllum* Hayasaka, 1924

Subgenus *Waagenophyllum* Hayasaka, 1924

Waagenophyllum (*Waagenophyllum*) *tachtabulasicum* Iljina, 1997, Plate 3, 3, 4

Waagenophyllum (*Waagenophyllum*) *tachtabulasicum:* Ilyina, 1997, p. 136, Plate 2, 1; Fig. 5, 4

Etymology. The species is named after the Takhtabulak Formation.

Holotype. PIN RAS, specimen 2376/344; the southeastern Pamir, Kur-Istyk River middle reaches, fourth left tributary, 1 km from a small lake; Upper Permian,

Plate 4.

(1, 2) *Waagenophyllum* (*Waagenophyllum*) *ramosum* Iljina, 1997.

(1) holotype 2376/316 PIN, thin section 3, transverse cross section of corallite (1a, b) ×5, and septum (1c) ×50; thin section 2, longitudinal cross sections of corallite (1d) ×5, and septum (1e) ×50, the Kur-Istyk River, fourth left tributary, Upper Permian, Dorashamian Stage, middle Takhtabulak Subformation. (2) specimen 2376/860 PIN, transverse cross sections, thin section 1, the Kur-Istyk River, left wall of the second left tributary, 1.5 km from the mouth, Upper Permian, Dorashamian Stage, middle Takhtabulak Subformation, ×5.

(3) *Iranophyllum* (*Iranophyllum*) *igrimjusicum* Iljina, 1997.

Holotype 2376/106 PIN, (3a) transverse cross section, thin section 2, ×2; (3b) longitudinal cross sections, thin section 1, ×3, 3.2 km from the Igrim'yuz River mouth, water-divide saddle of the ravine, Upper Permian, Murgabian Stage, Gana Formation.

(4, 5) *Waagenophyllum* (*Huayunophyllum*) *pamiricum* Iljina, 1997.

(4) holotype 2376/844 PIN, thin section 1, transverse cross sections of corallite (4a, b) ×10, wall and septum (4c) ×50; thin section 3, longitudinal cross sections of septum (4d) ×50, and corallite (4e) ×10, the Shur-Bulak River, left bank near the mouth, Upper Permian, Dorashamian Stage, middle Takhtabulak Subformation. (5) specimen 2376/142 PIN, transverse cross section, left bank of the Shindy River, 1.8 km from the mouth, Upper Permian, Dzhulfian Stage, lower Takhtabulak Subformation, ×7.

(6) *Lophophyllum* (*Lophbillidium*) *spinosum* (Martini, 1881).

Specimen 2376/525 PIN, (6a) transverse cross section, thin section 2; (6b) longitudinal cross section, thin section 3, the Mamazair-Bulak locality, 1 km east of the 467-km mark of the Pamirian road, Lower Permian, Bolorian Stage, lower Kochusu Subformation, ×6.

(7, 8) *Lophophyllum* (*Lophbillidium*) *martini* Schouppe et Stacul, 1955.

(7) specimen 2376/218 PIN, transverse cross section, thin section 7, ×4, the Mamazair-Bulak locality, 1 km east of the 467-km mark of the Pamirian road, Lower Permian, Bolorian Stage, lower Kochusu Subformation. (8) specimen 2376/345 PIN, transverse cross section, thin section 3, ×4, the same locality and stratigraphic position.

(9) *Wannerophyllum cristatum* (Gerth, 1921).

Specimen 2376/572 PIN, (9a) transverse cross section, thin section 8; (9b) longitudinal cross section, thin section 6, the Kur-Teke River upper reaches, second left tributary, Lower Permian, Bolorian Stage, lower Kochusu Subformation, ×5.

Dorashamian Stage, middle Takhtabulak Subformation.

Description. Corallites are cylindrical, 8.5–10 mm across. Septa, (24–25) × 2, are thick owing to stereoplasma accumulations and wedge-shaped, with sharpened axial ends (Plate 3, 3a, 4a). In length, the septum of the second order is equal to 2/3 or 3/4 of the septum of the first order. Septa are composed of one row of small (0.05–0.06 mm) simple trabeculae (Plate 3, 4b) inclined in the septum plane at the angle of 50°–75° (Plate 3, 4c). The dissepimentarium width slightly exceeds one third of the corallite diameter. The outer zone consists of two to three rows of small elongated dissepiments, whereas the inner zone includes one to two rows of large dissepiments (Plate 3, 3b). Lonsdaleoid dissepiments are locally developed. Tabulae (clinotabulae) are initially almost vertical and then become less inclined and horizontal or even approach the axial columella. The axial columella is rounded, occupies one third of the corallite diameter, and consists of pyramidal tabulae (5–7 per mm), the median lamina, which partly crosses the columella and commonly joins the counterseptum, and a few short radial laminae. The wall is thin (0.2–0.4 mm).

Comparison. The described species differs from *W.* (*W.*) *polyseptatum* Minato, 1995 and *W.* (*W.*) *orlensis* Kropatcheva in a smaller diameter and lesser number of septa, as well as in the presence of lonsdaleoid dissepiments. In addition, the wedge-shaped form of the septa differs it from *W.* (*W.*) *orlensis*.

Distribution. The Upper Permian, Dorashamian Stage, middle Takhtabulak Subformation.

Material. Two specimens from the Muzdu-Bulak locality, one specimen from the Dunkel'dyk locality, and one specimen from the Kur-Istyk locality.

Waagenophyllum (*Waagenophyllum*) *crassiseptum* Iljina, 1997, Plate 3, 5–7

Waagenophyllum (*Waagenophyllum*) *crassiseptum* Ilyina, 1997, p. 136, Plate 2, 3–5; Fig. 5, 1, 2

Etymology. The name of the species was derived from Latin "crassus" (thick) and "septum" (fence).

Holotype. PIN RAS, specimen 2376/770; the southeastern Pamir, Kutal' ravine, Zou-Tash locality, the height 3895.3 m; Upper Permian, Dorashamian Stage, middle Takhtabulak Subformation.

Description. Corallite diameter is 6–7 mm. The number of septa is (22–23) × 2 (Plate 3, 5, 6a). Axial ends of septa of the first order are thickened. Septa of the first order reach or almost reach the axial columella, retreating only in the calyx (Plate 3, 7a). Septa of the second order are long, equal to 3/4 to 9/10 of the first-order septa length. The septal microstructure is formed by a combination of simple (0.06–0.07 mm) and complex (0.024–0.28 mm) trabeculae. Most septa are completely built of complex trabeculae (Plate 3, 6c, 7b). Locally, in the middle part of septa, simple trabeculae are arranged in the one-row mode. The dissepimentarium consists of one to two rows of elongated steeply inclined dissepiments (Plate 3, 6b). Tabulae are represented by clinotabulae. The axial columella is rounded or, less commonly, oval; it is large, equal to one-third of the corallite diameter; axial tabulae are slightly convex, pyramidal, and densely arranged (7–8 per mm). The thin median lamina commonly crosses the entire columella. Radial laminae are numerous, often discrete. The wall is thin, 0.1–0.3 mm.

Comparison. The species differs from the closest *W.* (*W.*) *akasakense* (Yabe, 1902) and *W.* (*W.*) *compactum* Minato et Kato (1965) species in the septum form. In addition, it differs from *W.* (*W.*) *akasakense* in the absence of spheroidal dissepiments and development of the median lamina in the axial columella; as compared to *W.* (*W.*) *compactum*, it has lesser dimensions and a less dense arrangement of axial tabulae.

Distribution. The Upper Permian, Dorashamian Stage, middle Takhtabulak Subformation of the southeastern Pamirs.

Material. Seven specimens from the Kutal' locality and two specimens from the Ak-Bura locality.

Waagenophyllum (*Waagenophyllum*) *ramosum* Iljina, 1997, Plate 4, 1, 2

Waagenophyllum (*Waagenophyllum*) *ramosum* Iljina, 1997, p. 137, Plate 2, 6

Etymology. The name of the species is derived from "ramosum," which means "branching" in Latin.

Holotype. PIN RAS, specimen 2376/316: the southeastern Pamir, Kur-Istyk River, fourth left tributary; Upper Permian, Dorashamian Stage, middle Takhtabulak Subformation.

Description. Corallites are 6–9 mm across with the calyx up to 11 mm in diameter. Septa are thin, (24–26) × 2 in number, and have sharpened axial ends. Septa of the first order reach or almost reach the axial columella (Plate 4, 1a, b, 2). Septa of the second order are equal in length to 1/2 to 2/3 of the first-order septa. Septa are built of one row of small (0.05 mm) simple trabeculae (Plate 4, 1c) inclined in the septal plane at the angle of 65°–70° (Plate 4, 1e). The dissepimentarium occupies less than one-fourth of the corallite radius and consists of two to three rows of relatively large, elongated, strongly inclined dissepiments (Plate 4, 1d). Sometimes, one row of small semispherical dissepiments is sporadically developed along the periphery (Plate 4, 1c). Tabulae (clinotabulae) occupy one-third of the corallite radius. The axial columella is less than one-third of the corallite diameter; it includes the densely arranged (7–9 per mm) helmet-shaped axial tabulae, the thin median lamina, which crosses almost the entire columella, and numerous radial laminae. The wall is thin (0.1–0.1 mm).

Comparison. The species differs from *W.* (*W.*) *pulchrum* (Maeda, Hamada, 1962) in a more regular and more dense arrangement of tabulae, as well as in the

almost complete absence of small hemispherical dissepiments along the periphery (in *W.* (*W.*) *pulchrum*, they form more than two rows). It differs from *W. lui* Tseng, 1959 in a more dense arrangement of axial tabulae (in *W. lui*, their density is two per mm) and in shorter septa of the second order.

Distribution. The Upper Permian, Dzhulfian Stage, middle Takhtabulak Subformation of the southeastern Pamir.

Material. The Dzhulfian Stage: three specimens from the Igrim'yuz locality, one specimen from the Muzdu-Bulak locality, one specimen from the Shindy locality, one specimen from the Kur-Istyk locality, and two specimens from the Kastanat-Dzhilga locality. The Dorashamian Stage: six specimens from the Kur-Istyk locality and two specimens from the Kutal' locality.

Subgenus *Huayunophyllum* Tseng, 1959

Waagenophyllum (*Huayunophyllum*) *pamiricum* Iljina, 1997

Waagenophyllum (*Huayunophyllum*) *pamiricum* Ilyina, 1997, p. 137, Plate 2, 7

Etymology. The name of the species is derived from the geographical name of the area where it was first found.

Holotype. PIN RAS, specimen 2376/844; the southeastern Pamir, Shur-Bulak River; Upper Permian Dorashamian Stage, middle Takhtabulak Subformation.

Description. Corallites are small, from 2.3 to 3 mm, sometimes to 4.2 mm across. The number of septa is (12–16 to 20) × 2 (Plate 4, 4a, b, 5). Septa of the first order are long, reaching the axial columella, and thin in tabularium but thickened in the peripheral area owing to stereoplasma accumulation. In the last case, they jointly form the rim, which is particularly distinct in the calyx (Plate 4, 5). Septa of the second order are equal in length to 1/2 or 3/4 of the first-order septa. Septa are built of very small (0.03–0.04 mm) simple trabeculae (Plate 4, 4c) inclined in the septum plane at the angle of 75°–77° (Plate 4, 4d). The dissepiment zone occupies from 1/5–1/6 to 1/3 of the corallite diameter. It consists of one to two rows of elongated, almost vertical, dissepiments (Plate 4, 4e). Locally, there are small rounded dissepiments developed near the wall. The tabulae zone is narrow, equal to 1/11–1/8 of the corallite diameter. Tabulae are represented by steeply inclined clinotabulae. The axial columella is strongly variable within the single colony in both the dimensions and structure (Plate 4, 4a, b; 5). It occupies 1/3–1/2 of the corallite diameter. The median lamina is usually distinct and crosses the entire axial columella; sometimes, it is missing. Radial laminae are rare or missing; in this case, an auloslike structure appears. The axial tabulae are hemispherical or irregular in form; they are flat or vertically elongated in the area where the median lamina appears. Boundaries of the axial columella are either distinct or blurred. The structure and dimensions of the axial columella change in the interval of 2 mm (Plate 4, 4e). The wall is thick.

Comparison. The considered species differs from *W.* (*H.*) *kanense* (Kuwano, 1959) in lesser dimensions of corallites having the well-developed septa of the second order. It also differs from *W.* (*H.*) *teratoideum* (Fontaine, 1961) in a lesser number of septa in specimens with equal diameters and also in development of the median lamina.

Distribution. The Upper Permian, Dzhulfian Stage, lower Takhtabulak Subformation and Dorashamian Stage, middle Takhtabulak Subformation of the southeastern Pamir.

Material. The Dzhulfian Stage: two specimens from the Igrim'yuz locality, two specimens from the Kur-Istyk locality, five specimens from the Shindy locality. Dorashamian Stage, two specimens from the Kur-Istyk locality, two specimens from the Karabeles locality, one specimen from the Shur-Bulak locality, and two specimens from the Ak-Bura locality.

Subfamily Wentzellinae Hudson, 1958

Genus *Iranophyllum* Douglas, 1936

Subgenus *Iranophyllum* Douglas, 1936

Iranophyllum (*Iranophyllum*) *igrimjusicum* Iljina, 1997, Plate 4, 3

Iranophyllum (*Iranophyllum*) *igrimjusicum* Ilyina, 1997, p. 137, Plate 2, 2

Etymology. The species is named after the Igrim'yuz River.

Holotype. PIN RAS, specimen 2376/106. The Southeastern Pamir, Vostochnyi Igrim'yuz Creek, 3.2 km from the Igrim'yuz River mouth, watershed saddle; Upper Permian, Murgabian Stage, Gana Formation.

Description. Corals are solitary, wide-conic, 4.5 cm high, and 35–40 mm across. Septa are of four orders. Septa of the first order (23–24) are thin and long, with carina-shaped growths; they thicken in the median part and reach the axial columella. When compared to them, septa of the second order are nine-tenths as long, septa of the third order are three-fourths as long, and those of the fourth order are two-thirds as long (Plate, 4, 3a). Septa are built of small (0.07 mm) simple trabeculae arranged in the septum plane at the angle of 75°–80°. Dissepiments are small, gently inclined toward the center, and occupy a wide zone along the coral periphery (one-fourth of its diameter). Locally, small lonsdaleoid dissepiments are developed. Tabulae are thin and gently inclined toward the axial columella; they occupy a narrow zone (one-ninth of the diameter) and locally are complicated by supplementary laminae (Plate 4, 3b). The axial columella is equal to one-tenth of the coral diameter (Plate 4, 3a) and consists of the thin median lamina that crosses the columella, numerous radial laminae, and axial tabulae. The last elements are pyra-

midal in form, strongly elevated, and densely arranged (six to seven per mm). The wall is thin.

Comparison. The considered species differs from *I. darvasicum* Pyzhjanov, 1966, in that the axial columella is twice as narrow, in denser arrangement of axial tabulae, and in development of septa of the third and fourth orders (*I. darvasicum*, probably, also has septa of the third order).

Distribution. The Upper Permian, Murgabian Stage, Gana Formation of the southeastern Pamir.

Material. Two specimens from the Vostochnyi Igrim'yuz valley.

CONCLUSION

(1) Permian rugose corals of the southeastern Pamir are first identified and divided into stage, formation, and subformation coral assemblages. In combination with other faunal groups, these assemblages help to date the stratigraphic levels. Three zones of rugose corals are established.

(2) Rugosa assemblages are correlated with the assemblages of guide faunal groups.

(3) Correlation with other regions of the Tethyan realm demonstrated the faunal exchange of the southeastern Pamir basin with the Indonesia (Timor) basin in the Bolorian, with the Iranian and Caracorum basins in the Murgabian–Midian, and with the Himalayas and South China basins in the Dzulfian and Dorashamian periods.

(4) The increased population of endemics against the background of a general decrease in coral abundance and diversity during the Dorashamian suggests isolation of the considered basin at that time.

(5) The absence of solitary pleurophyllids characteristic of some other Tethyan regions in the Dorashamian time is probably caused by unfavorable ecologic environments.

(6) The Dorashamian waagenophyllids show a complicated microstructure. The extinction of these corals at the Permian–Triassic boundary was probably caused by their specialization.

(7) The thorough study of the microstructure allowed the coral taxonomy to be revised. Based on the difference in the trabeculae arrangement in the septal plane, *Verbeekiella* and *Wannerophyllum* are classified again as independent genera. In addition, several families are united: Timorphyllidae and Verbeekiellidae (junior synonym), both displaying the fan-shaped trabeculae arrangement; and Lophophyllidae and Wannerophyllidae (junior synonym), both having the monoclinal trabeculae arrangement. The analysis of the microstructure in various rugose corals revealed its taxonomic significance.

(8) Based on morphologic and ontogenetic peculiarities, new genera *Pseudoverbeekiella* and *Timorocarinophyllum* and ten new species are identified and described.

Reviewers V.Yu. Dmitriev and B.S. Sokolov

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Differentiation of Foraminiferal Assemblages and Dynamics of Their Taxonomic Diversification in the Jurassic

K. I. Kuznetsova

Geological Institute, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia

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Abstract—Data on Jurassic foraminifers from key localities in continents and oceans is analyzed. Their distribution range spans the stratigraphic interval from the Aalenian to the Tithonian equal to 32 million years. Among 23 localities (including 7 Jurassic sections recovered by deep sea drilling), data for 9 are originally obtained. Information on other localities is presented on the basis of published data. The assemblages studied comprise three types of foraminiferal fauna corresponding to those distinguished by V.A. Basov for different climatic zones: Arctic (cold), Boreal (moderately cold), and Tethyan (warm tropical to subtropical). It is established that faunas different not only in dominant components but also in taxonomic composition of a high rank were characteristic of the Tethyan and Peritethyan regions in the Jurassic. Assemblages of different types are present in adjacent sections and even in the same section. Neither climatic fluctuations nor penetration of cold or warm water masses, as well as fauna migration and invasion, can explain this phenomenon. The mosaic distribution of faunal assemblages and alternation of different fauna types are the result of bathymetric differentiation and related environmental factors such as temperature, type of bottom substrate, hydrodynamics, and distance to the shore line.

Key words: *foraminifers, assemblages, differentiation, paleobiogeography, endemic forms, cosmopolitan forms, bathymetric zonation.*

Any stratigraphic and paleobiogeographic study at the stage of generalization implies systematization of data on regularities in the evolution and distribution of the fauna and on factors influencing these subjects.

The Jurassic is of great interest in this aspect because of major reorganizations in the geologic history of continents and oceans that happened at that time.

The Jurassic represents the middle part of the Mesozoic; therefore, its fauna retains the peculiar Early Mesozoic (Triassic) features and, accordingly, genetic relations of many groups of organisms with those of preceding epochs. At the same time, it was an epoch of formation of new groups, whose evolution and further development were related to the Cretaceous. All the factors left their imprints on foraminifers, benthic assemblages of which inhabited areas from the Arctic regions to the southern hemisphere and were unique in abundance and diversity during the Mesozoic.

Important information on the Jurassic foraminifers of continents and oceans has been acquired to date. Deep sea drilling supplied us with material that enhanced our understanding of this fauna. Even a general study of the published data shows the marked diversity of the Jurassic foraminiferal assemblages, the variability of their taxonomic composition and structure, and mosaic patterns of their distribution in the Jurassic sedimentary basins (Fig. 1).

Analysis of the composition and data on stratigraphic and paleobiogeographic ranges of the Jurassic foraminifers (Table 1) shows the following.

The foraminiferal assemblages of the studied stratigraphic interval (the Aalenian–Tithonian) are represented exclusively by benthic forms. Planktonic foraminifers (*Eoglobigerina, Globuligerina,* and *Conoglobigerina* genera) are rare and represented by 11 species only. These are known from the early Bajocian to the Tithonian inclusive and are confined to the Tethyan, Peritethyan, and Subboreal zones. Planktonic foraminifers were not found in the Arctic regions (Arctic Canada, northern Siberia, and Arctic islands). In some localities (Turkey, the Russian platform, and Crimea), populations of planktonic foraminifers are as abundant as thousands of shells per 50 g of rock. Usually, they are sporadic and represented by single specimens. The foraminiferal assemblages studied comprise 177 genera, which belong to 16 orders. The number of genera in each order is different. More abundant are genera of the Lituolida, Ataxophragmiida, Lagenida, and Rotaliida orders. The dynamics of generic diversification in the Lagenida and Rotaliida orders during the Jurassic is shown in Fig. 2. The comparison of stratigraphic and paleobiogeographic ranges of genera showed that cosmopolitan forms existed longer, as compared to endemic (both Tethyan and Boreal) genera demonstrating narrow stratigraphic intervals rarely exceeding a stage (geologic age).

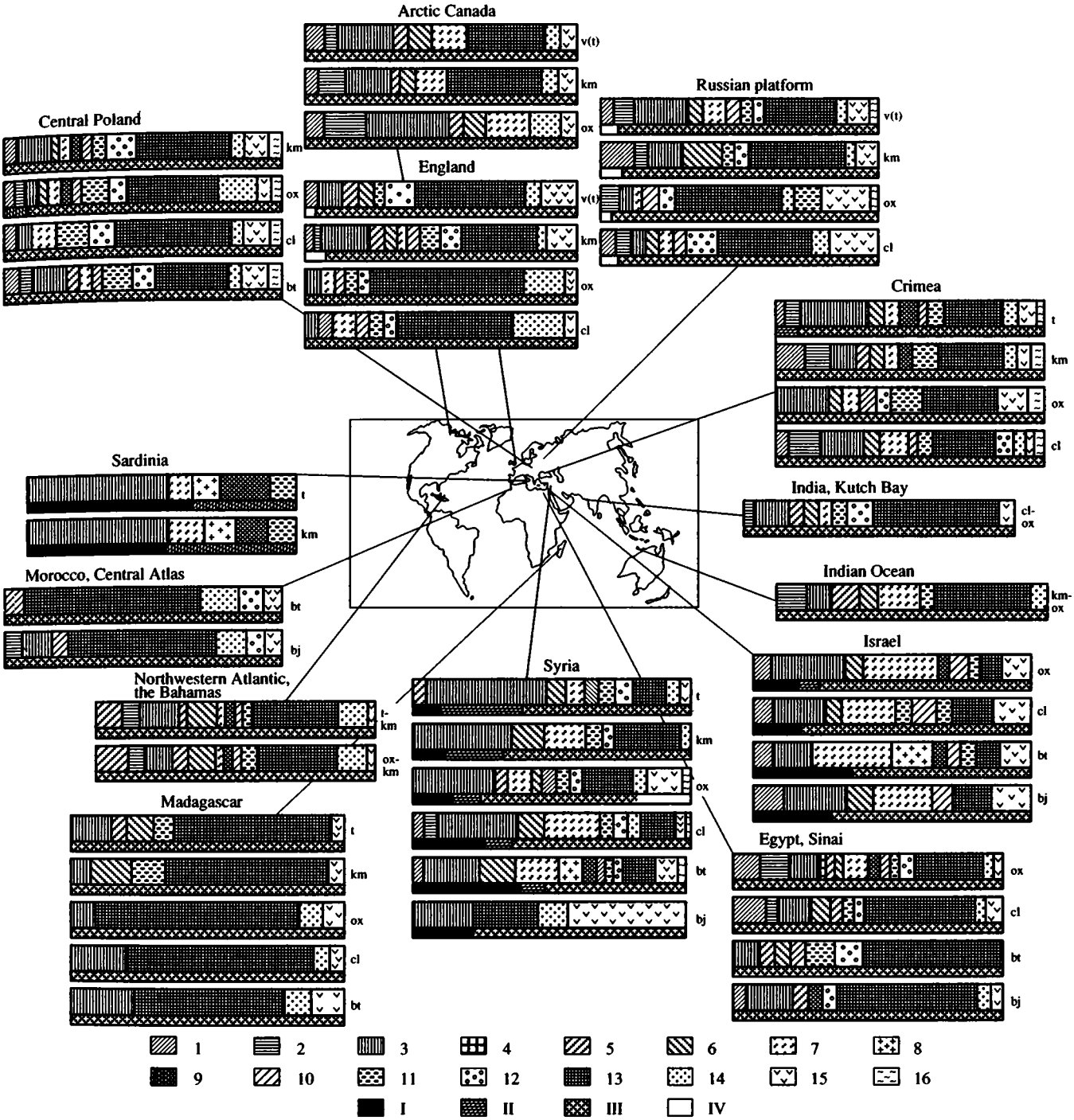
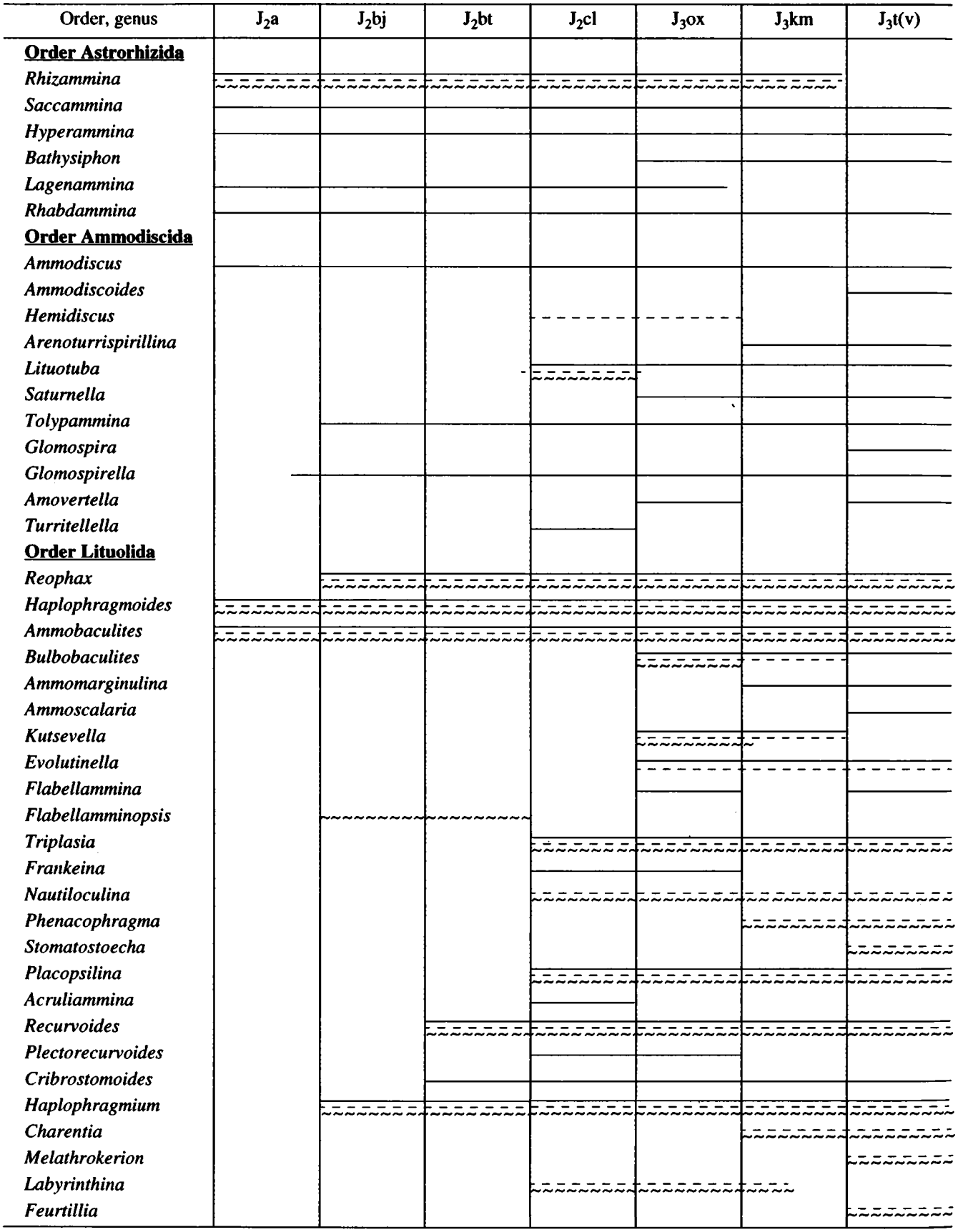


Fig. 1. Taxonomic composition of Jurassic foraminiferal assemblages (by orders, %): Madagascar after Espitalié and Sigal (1963) and Kuznetsova (1976); Syria after Kuznetsova *et al.*, (1996); Israel after Mayne (1966) and Picard and Hirsch (1987); Egypt, Sinai, after Said and Barakat (1958); Morocco, Central Atlas, after Sossipatrova and Rahhali (1974); Riegraf *et al.* (1984); India, Kutch Bay, after Bhalla and Abbas (1978); Indian Ocean, Argo Abyssal Plain, DSDP Leg 27, Site 261, after Kuznetsova (1974); Crimea after Kuznetsova and Gorbachik (1985); Sardinia after Azema *et al.* (1977); Poland after Bielecka (1960), Bielecka and Styk (1968, 1981), and Bielecka and Kuznetsova (1969); England after Cordey (1963) and Kuznetsova (1979); Russian platform after Kuznetsova (1979), Grigyalis (1985), and Azbel' (1989); Arctic Canada after Souaya (1976) and Hedinger (1995); northwestern Atlantic, the Bahamas, DSDP sites 100, 101, 534 after Luterbacher (1972), Gradstein (1978, 1983), and Gradstein and Sheridan (1983). Orders of foraminifers: (1) Astrorhizida, (2) Ammodiscida, (3) Lituolida, (4) Rzehakinida, (5) Trochamminida, (6) Textulariida, (7) Ataxophragmiida, (8) Orbitolinida, (9) Involutinida, (10) Cornuspirida, (11) Spirillinida, (12) Miliolida, (13) Lagenida, (14) Polymorphinida, (15) Rotaliida, (16) Globigerinida; (I) Tethyan endemics; (II) Tethyan and Subtethyan forms; (III) Boreal and cosmopolitan forms; (IV) Boreal endemics.

Table 1. Stratigraphic and paleogeographic ranges of the Jurassic foraminifers



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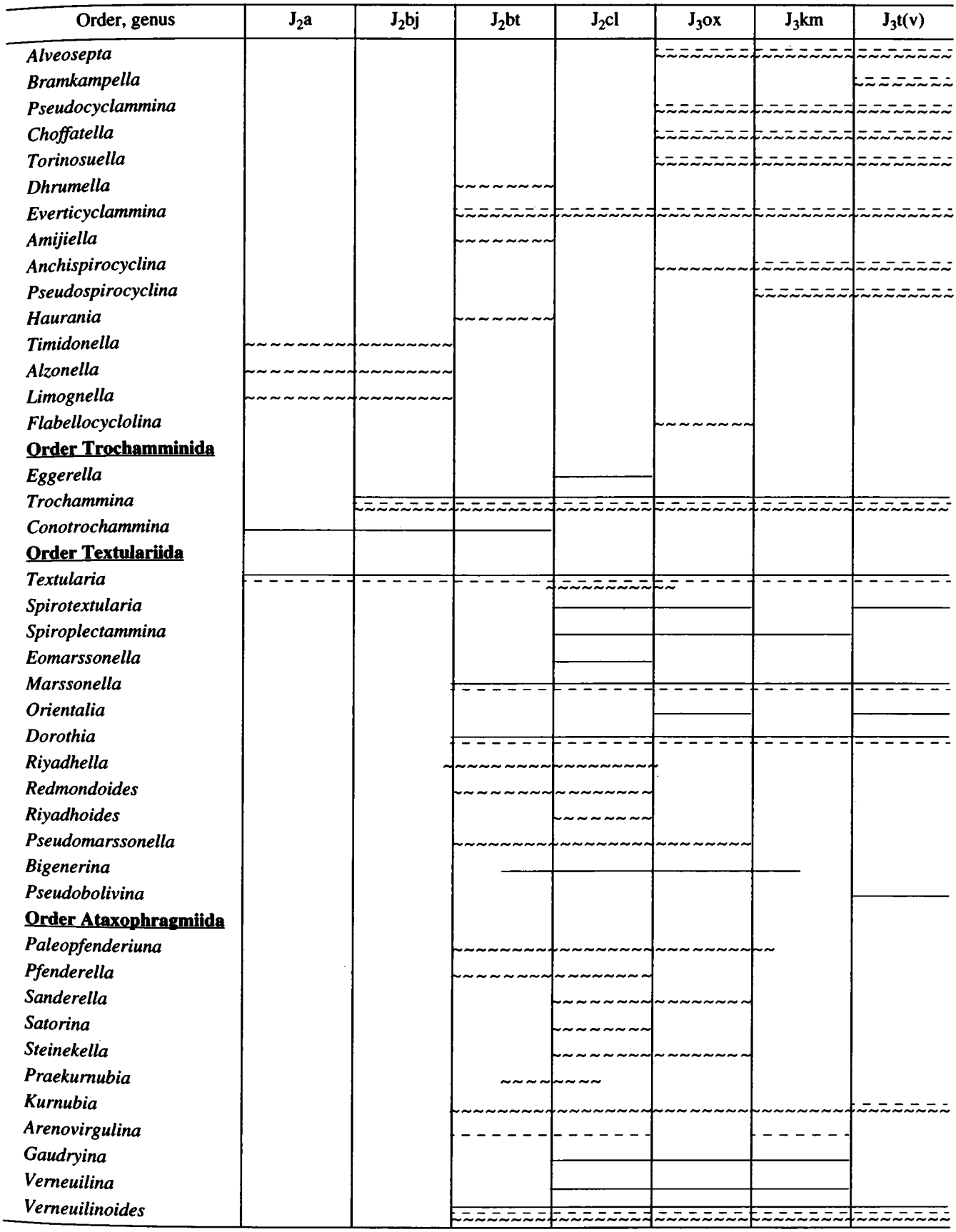

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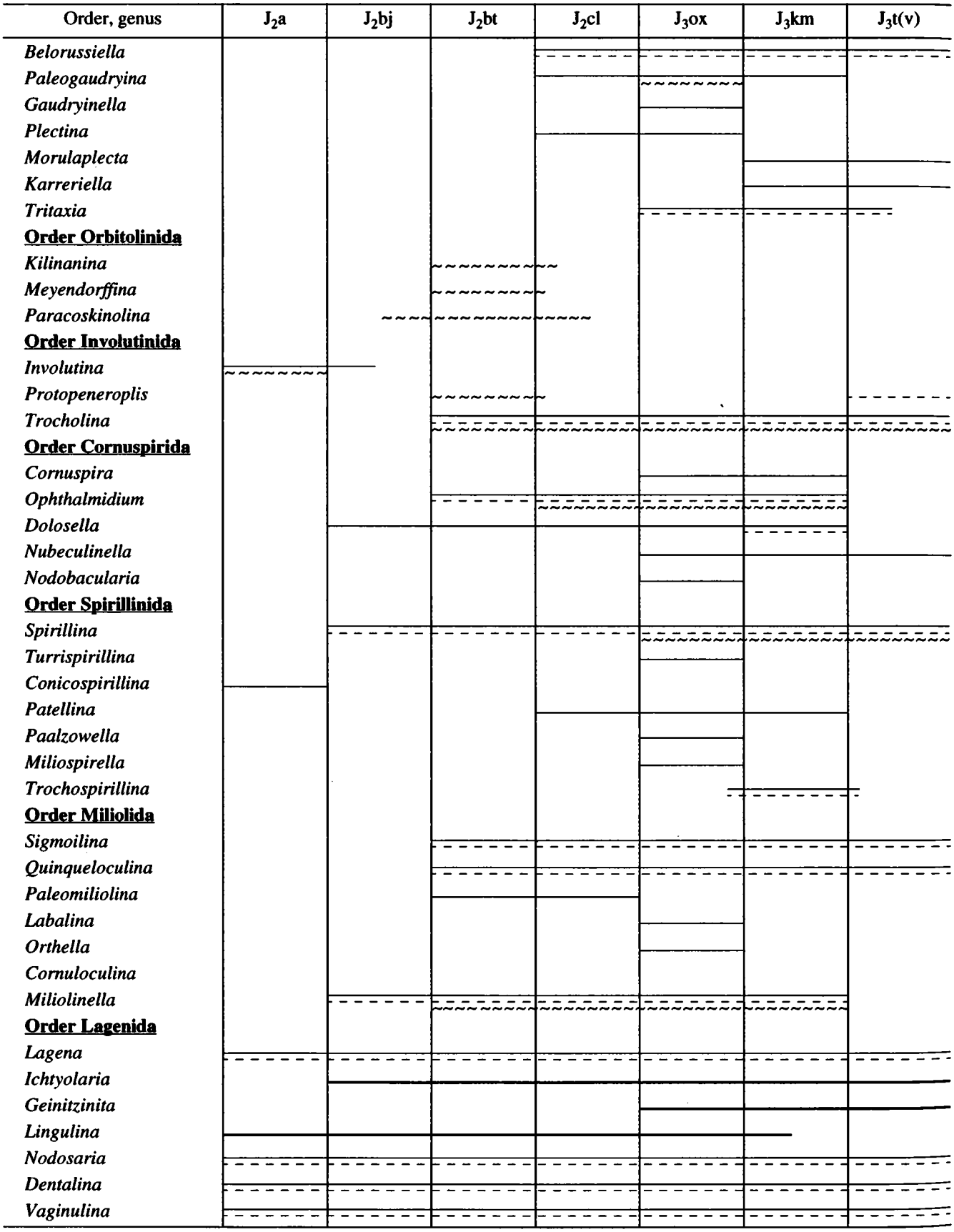

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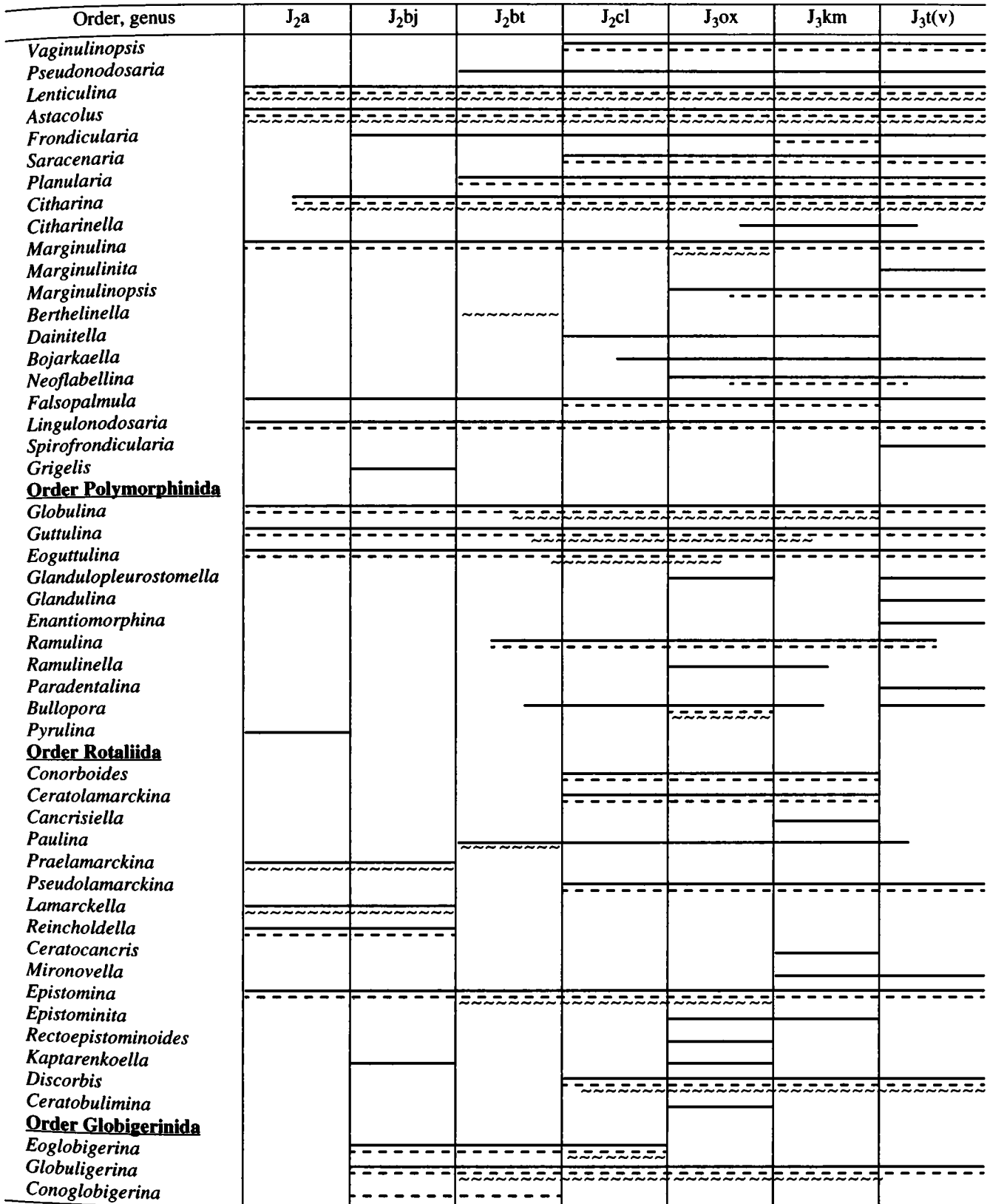

——— – Distribution ranges of foraminifers in the Boreal and Arctic areas.

- - - - – Same in the Peritethys.

~~~~~ – Same in the Tethys.
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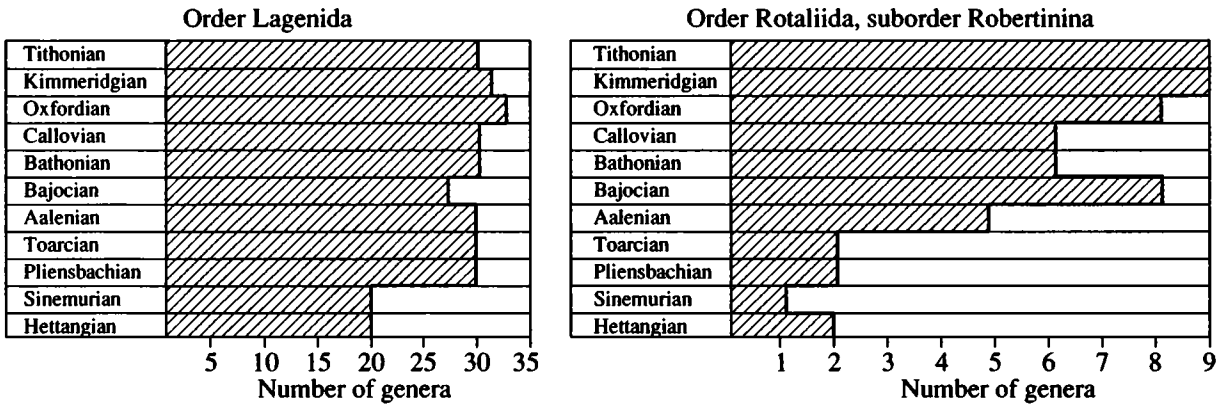

Fig. 2. Number of Jurassic genera in orders Lagenida and Rotaliida (suborder Robertinina), after Tappan and Loeblich, (1988).

Gordon (1970) was first in systematization of available data and distinguished two types of foraminiferal assemblages: shelf communities subdivided into three subtypes, and Tethyan assemblages, which include two groups. Gordon's subtypes of the shelf assemblages are as follows: (1) the nodosariid and mixed nodosariid subtype, (2) the subtype dominated by primitive agglutinated forms, and (3) the subtype of secretory benthic foraminifers different from nodosariids of the first subtype. The Tethyan foraminifers were subdivided by Gordon into assemblages with dominant agglutinated and planktonic forms.

V.A. Basov (1974) suggested the typification of foraminiferal faunas on the basis of the areal–genetic principle. He distinguished three types of fauna corresponding to various climatic zones: (1) the nodosariid–ammodiscid found in the Arctic zone; (2) the nodosariid–epistominid found in the Boreal and Subboreal moderately warm zones; and (3) the cyclaminid–pavonitinid (later renamed as cyclaminid–pfenderinid) fauna of tropical and subequatorial basins characteristic of the Tethys. The typification clarified our understanding of the diversity of foraminifer assemblages and elucidated their paleobiogeographic features. Nevertheless, the subsequent accumulation of data and especially the detailed study on foraminifers from the Jurassic sections of the Tethys (northern Africa, southwestern Europe, and countries of the Eastern Mediterranean region) showed that this typification was not consistently successful as was expected. It was found that at least the cyclaminid-pfendirinid (Tethyan) and nodosariid–epistominid (Boreal) faunas inhabited the same climatic zone of the subequatorial Tethys in the Jurassic time. Geographic ranges of these dissimilar assemblages were contiguous, and changes in the composition of the assemblages occurred quickly in the lateral direction, since it was established in sections spaced only a few dozens of kilometres apart. Such abrupt changes in the composition of bottom assemblages are exemplified by the Bathonian assemblages from the Anti-Lebanon and Maritime Ranges of Syria; the Callovian and Oxfordian assemblages from the Maritime Range of western Syria (Kuznetsova *et al.*, 1996); the Bajocian and Bathonian assemblages of Morocco (Hottinger, 1967; Sossipatrova and Rohhali, 1977); the Callovian, Oxfordian, and Kimmeridgian assemblages of Israel and Egypt (Maync, 1966; Said and Barakat, 1958); the Tithonian assemblages from the mountains and eastern area of Crimea (Gorbachik and Kuznetsova, 1994), and also from Portugal (Stam, 1986; Exton and Gradstein, 1984) and Spain (Fourcade *et al.*, 1977). Moreover, I observed in many sections that the Tethyan endemic assemblages of the cyclaminid–pfenderinid type confined to carbonate rocks alternate with assemblages of the nodosariid–epistominid type occurring in terrigenous and terrigenous–carbonate interlayers. It is notable that this alternation may have taken place within one stage or sometimes within one foraminiferal zone.

The examples cited above show that Basov's typification (1974), quite adequate as a whole, ignored an important additional factor that influenced the evolution and distribution of benthic faunas in the Jurassic seas. I believe that the bathymetric zoning of basins, important for the distribution of benthic organisms, could be such a factor. Later, Basov also demonstrated its importance for the distribution of Jurassic and Cretaceous foraminiferal assemblages, and, when analyzing their changes in various facies of the Mesozoic oceanic sediments, he showed that the composition of assemblages was dependent on the depth of their dwelling (Basov and Vasilenko, 1986; Basov, 1991).

My analysis of materials on the problem shows that the available data characterize Jurassic foraminifers from all bathymetric zones from the abyssal to littoral settings of past basins. These data are supplemented with information on the facies confinement of benthic organisms and on their taxonomic composition in all climatic zones of the Jurassic time. It should be pointed out that the climatic zones can be easily recognized by the geographic ranges of the faunas studied, whereas their bathymetric confinement is detectable on the basis

Table 2. Data on Jurassic deposits penetrated by deep sea drilling holes

Region	DSDP sites	Age of deposits (by foraminifers)	Depth (m) at present	Thickness of deposits (m)	Basin depth in the Jurassic (km)	References
North Atlantic	99	Oxfordian–Kimmeridgian	4914		Less than 3	Luterbacher, 1972
North Atlantic	100	Oxfordian–Kimmeridgian–Tithonian	5325		3	Luterbacher, 1972
North Atlantic	105	Oxfordian–Kimmeridgian–Tithonian	5251	630	3	Luterbacher, 1972
North Atlantic	391C	Tithonian	4963	1412	3–3.5	Gradstein, 1978
North Atlantic	534A	Middle Callovian–Tithonian	4974	1636	3–3.5	Gradstein, 1983
North Atlantic	367	Oxfordian–Kimmeridgian–Tithonian	4758		3	Kuznetsova and Seibold, 1978
North Atlantic	416	Tithonian			3–3.5	Sliter, 1980
Indian Ocean	216	Oxfordian–Kimmeridgian	5125	542	3	Kuznetsova, 1974; Bartenstein, 1974

of some features often missed by investigators, because it is difficult to recognize them in sections. As a result, we observe similar assemblages from shallow-water shelf areas of the Arctic and Boreal regions, on the one hand, and from deeper zones of tropical seas, on the other.

Naturally, other factors, such as the near-bottom temperature, the composition and character of sediments (the substrate for living organisms), the distance to the shoreline and related hydrodynamic regime, the chemistry and salinity of the water, the content of calcium carbonate in it, and the trophic and other conditions, are directly related to the bathymetric zoning of a basin.

It is difficult to present a detailed analysis of all biotic and abiotic factors in a short paper; therefore, I shall try to summarize briefly only data on principal reasons including the depth and bathymetric zones of a basin, as well as the facies type of deposits, since they all control the taxonomic composition of foraminiferal assemblages, their structure, and geographic ranges.

BATHYMETRIC ZONING AND COMPOSITION OF FORAMINIFERAL ASSEMBLAGES

Abyssal assemblages are established in many sections of the Jurassic deposits penetrated by deep sea drilling (Fig. 1) in the northwestern (Sites 100, 101, 105, 534) and eastern Atlantic Ocean (Sites 370, 367, 416), and also in the Indian Ocean (Sites 216, 261). Data on the Jurassic foraminifers are presented below in Table 2 modified after Gradstein (1983).

The abyssal assemblages, studied in the cited works, comprise representatives of 13 orders and about 60 genera, including forms with secretory calcite shells from the Nodosariidae, Lenticulinidae, Ceratobuliminidae, and Spirillinidae families. They indicate that sediments and coeval foraminifers were formed above the carbonate compensation depth (CCD). It is also remarkable that Luterbacher (1972) and Gradstein (1983) reported about planktonic foraminifers of the *Globuligerina* genus in the Jurassic deposits of the abyssal bathymetric zone. The available data shows that deposits directly overlying the basaltic basement are the Late Jurassic (Oxfordian, Kimmeridgian, and Tithonian) in age. The only exception is Site 534A, where the recovered deposits are referred to the interval from the Callovian to the Tithonian (Gradstein, 1983). It should be pointed out that abyssal assemblages of foraminifers are present only in oceanic sediments penetrated by deep sea drilling. Jurassic deposits of this bathymetric zone have not been found in continents. Deposits of the bathyal zone more widespread in Jurassic sea basins are known in a section of the northwestern Atlantic Ocean (the Bahamas region), western Europe (Holzer, 1969; Feninger and Holzer, 1970), and Arctric Canada (Hedinger, 1993).

Along with primitive agglutinated forms from the Astrorhizida order, the bathyal foraminiferal assemblages comprise ammodiscids, lituolids, and trochamminids. Foraminifers with a secretory shell (lagenids and epistominids) are less abundant. Planktonic foraminifers have not been found. Hedinger (1993) reported on the characteristic composition of foraminiferal assemblages from the Upper Jurassic deposits (the Oxfordian–Volgian) studied in the Richardson Mountains (the Aklavik Range of Arctic Canada). The assemblage from clay and shale beds of the Husky Formation includes here 95 species of agglutinated foraminifers belonging to 29 genera, as well as 46 secretory species of 20 genera. The density of these populations is as high as more than 100 shells per g of the rock sample.

The foraminiferal assemblages from the Middle Jurassic sections of the Central Atlas (Morocco) presumably characterized the bathyal and partly epibathyal deposits (Sossipatrova and Rohhali, 1977) (Fig. 1). Lagenids comprising the diverse and abundant *Citharina, Lenticulina, Astacolus,* and *Planularia* species are

the dominant group. Similar but more diverse assemblages, which include lituolids, trochamminids, and ataxophragmiids, are described from the Middle Jurassic (Aalenian–Callovian) deposits of France (Wernli, 1971), the Middle and Upper Jurassic deposits of Germany (Lutze, 1960), and from the eastern Atlantic Ocean (Riegraf *et al.*, 1984; Sliter, 1980).

The transition to the epibathyal zone is marked by an enrichment and transformation of foraminiferal assemblages, in which epistominids and ceratobuliminids (the Rotaliida order), coexisting with diverse agglutinated and secretory foraminifers, play a leading role (Fig. 1). The geographic range of these assemblages is wider than that of the bathyal assemblages. They are known from the Middle and Upper Jurassic (Callovian–Tithonian) deposits of Portugal (Stam, 1986), the Oxfordian deposits of France (Bastien and Sigal, 1962), and the Middle and Upper Jurassic (Bathonian–Tithonian) deposits of Madagascar (Espitalié et Sigal, 1963), southern England (Cordey, 1963; Lloyd, 1959, 1962), central Poland (Bielecka, 1960; Bielecka and Styk, 1981), eastern Crimea (Kuznetsova and Gorbachik, 1985), Sinai (Said and Barakat, 1958), and the Anti-Lebanon Ridge in Syria (Kuznetsova, 1994; Kuznetsova and Dobrova, 1995; Kuznetsova *et al.*, 1996). The presence of *Globuligerina bathoniana* (Pazdro), *G. oxfordiana* (Grig.), and *G. calloviensis* Kuzn. is characteristic of sections in Syria, Crimea, Poland, and Portugal.

The further taxonomic diversification of foraminiferal assemblages is observed in the transition to the shallower areas of sea basins representing their sublittoral zone. The geographic differentiation of benthic assemblages from this zone becomes more pronounced. However, nodosariids (diverse ornamented and psilate *Lenticulina, Marginulina, Marginulinopis, Citharina*, and *Citharinella* genera) and diverse genera and species of epistominids are still the dominant groups. Forms with agglutinated shells such as large *Ammobaculites, Haplophragmoides, Evolitinella, Kutsevella* and other genera of lituolids preserve their significance. Abundant polymorphinids inhabited areas remote from the shoreline, where the hydrodynamic regime was less active. Benthic foraminifers of the sublittoral zone are clearly differentiated in accordance with latitudinal and climatic zones. In the Boreal areas, they are diverse and have the above-mentioned taxonomic composition, but, despite a good preservation, they rarely occur in abundance. The sublittoral and littoral deposits are known on the Russian platform, where in the lectostratotype of the Volgian Stage there is the bathymetric succession of the bathyal (?) (the upper Kimmeridgian–lower Volgian), epibathyal (the middle Volgian), and sublittoral to littoral (the upper Volgian) sediments. The deposits of the littoral zone are represented by shallow-water sediments, namely, by quartz, glauconite, and polymictic sands and sandstones, in places coarse-grained, which show numerous traces of erosional unconformities. These sediments bear a peculiar foraminiferal assemblage exclusively consisting of nodosariids, mainly of *Marginulina* species (Kuznetsova, 1979, Fig. 22).

Foraminiferal assemblages quite different in taxonomic composition inhabited shallow-water marine sediments of the Tethys. Cyclaminid–pfenderinid assemblages populated shallow and warm zones of sea basins rich in calcium carbonate. Such assemblages of larger foraminifers are unknown beyond the limits of these bathymetric zones of the Tethys. The assemblages include genera *Paleopfenderina, Meyendorffina, Kilianina, Timidonella, Alzonella, Paracoskinolina, Dhrumella, Paraurgonina, Haurania, Kurnubia, Redmondoides, Riyadhella, Bramkampella* and others. Representatives of nodosariids, epistominids, polymorphinids, and other smaller foraminifers either are not present or are represented by single shells seldomly encountered. Many genera and families of these assemblages are endemic. The species diversity is very low, whereas the density of populations for some species is very high (up to 1000 specimens per 100 g of dry rock). This is especially typical of genera *Kurnubia, Pseudocyclammina, Choffatella,* and *Anchispirocyclina*. Assemblages of this composition are present in the Bathonian, Callovian, Oxfordian (very locally), Kimmeridgian, and Tithonian deposits of Syria (Kuznetsova *et al.*, 1996), France (Peybernes, 1976), the eastern Carpathians (Dragastan, 1975), Sardinia (Azema *et al.*, 1977), Antalia (Bassoullet and Poisson, 1975), and Morocco (Hottinger, 1967). Incomplete assemblages consisting of *Anchispirocyclina, Choffatella,* and *Bramkampella* forms are found in the Tithonian carbonate rocks of the Crimea Mountains (Gorbachik and Kuznetsova, 1994) (Fig. 1). A brief review of the taxonomic composition of foraminiferal assemblages and their distribution in bathymetric zones of the Jurassic sea basins shows that the Jurassic benthic assemblages inhabited the whole series of settings from the abyssal to the littoral zones. Geographic ranges of Jurassic abyssal assemblages were limited.

More diverse assemblages, which include more than 60 genera of agglutinated and secretory foraminifers, are present in sediments from the depth of 2000–3500 m and have wider geographic ranges (the northwestern and central Atlantic and Indian Ocean). They are known only in oceanic sections, where the Upper Jurassic sediments directly overlie the basaltic basement.

The bathyal zone shows again the increasing taxonomic diversity of foraminiferal assemblages comprising here more than 80 genera with secretory and agglutinated shells, and also planktonic foraminifers appeared for the first time. The geographic range of assemblages from this zone is also widened, and they are known in northern, western, and central Europe; in the northern Russian platform; northern Siberia; Arctic islands; northwestern Africa; Hindustan; and the eastern Atlantic Ocean. The geographic differentiation of their taxonomic composition is poorly pronounced.

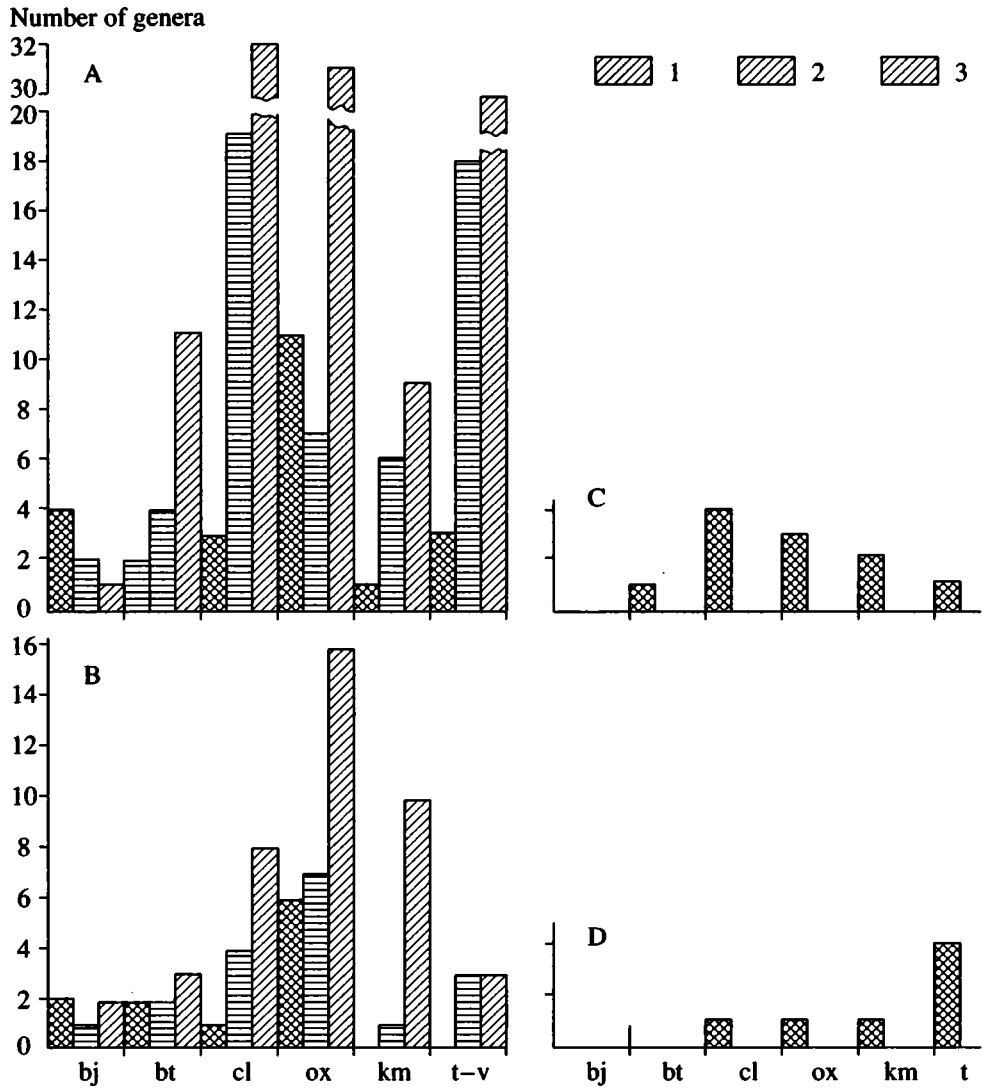

Fig. 3. Dynamics of species diversity in dominant genera of the Boreal and Tethyan fauna types of foraminifers: (1) Syria; (2) Crimea; (3) Russian platform and England. (A) *Lenticulina* and (B) *Epistomina* genera of the nodosariid–epistominid fauna type. (C) *Kurnubia* and (D) *Pseudocyclammina* genera of the cyclamminid–pfenderinid fauna type.

The epibathyal and outer sublittoral zones (500–200 m deep) display the presence of abundant and diverse foraminiferal assemblages, which inhabited wide areas in the Boreal and Tethyan provinces during the Jurassic. The number of genera, which belong to 16 orders, increases up to 100–110. The geographic differentiation is distinct.

The middle and inner sublittoral (200–100 m deep) and littoral zones comprise foraminiferal assemblages diverse in composition and including more than 100 genera. The geographic and climatic differentiation is extremely high. The number of endemic genera and families increases in the Tethys, whereas the diversity of endemic species grows in the Boreal zone. Geographic ranges of these assemblages are limited.

Changes in the taxonomic composition of foraminiferal assemblages in different bathymetric zones of the Jurassic seas show that, in addition to the long-lived cosmopolitan genera demonstrating a high degree of tolerance and wide geographic ranges, the assemblages in question also comprise endemic genera of highly specialized forms with a lower degree of tolerance, more limited geographic ranges, and, as a rule, shorter existence periods. Other peculiar features of genera of the two aforementioned types—endemic and cosmopolitan—are interesting as well. The quantitative analysis of generic and specific diversity is important, first of all for those genera that dominate in the different-type assemblages, i.e., in the Boreal nodosariid–epistominid (widespread) and Tethyan cyclaminid–pfenderinid (endemic) faunas (Fig. 3).

When the number of species of each indicated genera is compared, a clear regularity becomes obvious. The *Lenticulina* genus is present in deposits of all stages in territories under comparison, although the number of species of the genus in Syria, the Crimea, and the Russian platform differs greatly (Fig. 3A). For instance, the number of *Lenticulina* species in the Bathonian deposits is 2 in Syria, 4 in the Crimea, and 11 in the Russian platform. The distinction is even sharper for the Callovian and Oxfordian deposits. The Callovian deposits of the *Lenticulina* genus comprise 3 species in Syria, 19 in the Crimea, and 32 in the Russian platform. The Oxfordian deposits of Syria are known to include 11 species of the *Lenticulina* genus, while 7 are reported from the Crimea and 31 from the Russian platform. Some decrease in the species diversity of the *Lenticulina* genus is observed in the Kimmeridgian, but, in the Tithonian (Volgian), the number of species of the genus increases abruptly: the deposits of this age in the Crimea and Russian platform comprise 18 and 30 species, respectively, and they yielded only 4 species in Syria. A similar distribution trend is characteristic of *Epistomina* species, the second group dominant in the assemblages under consideration (Fig. 3B).

In other words, when cosmopolitan genera prevail in the Jurassic assemblges, the species diversity in the Jurassic basins of Syria is less than that in the Peritethyan basins of the same age and, especially, in the Russian platform and England, where these forms, diverse and abundant, are most widespread and dominant. Cosmopolitan genera, which have the widest (subglobal) geographic ranges, have long been existing, and many of them are known from the Early Mesozoic to the present day. A high species diversity is another characteristic feature of these fauna types that distinguishes them from the Tethyan endemic genera. The latter, exhibiting a low tolerance, high specialization, intricate shell structure, and limited geographic ranges, have narrow stratigraphic ranges and a low species diversity. Among 22 endemic genera that lived in the Jurassic tropical seas, more than half of them are represented by one species (genera *Dhrumella, Haurania, Amijiella, Meyendorffina, Timidonella, Alzonella, Flabellacyclolina,* and *Paracoskinolina*) or by 2 to 3 species (genera *Pseudocyclammina* and *Paleopfenderina*), and only some of them *(Kurnubia, Redmondoides,* and *Riyadhella*) are more diverse in species composition (Fig. 3C, 3D).

Hence, the analysis of the generic composition of the Jurassic foraminiferal assemblages and dynamics of their species diversity showed that nodosariids and epistominids had the highest species diversity and highest degree of tolerance among other benthic foraminifers in the Jurassic period of their evolution. Many genera of these families comprise species with narrow stratigraphic ranges, which is very important for the detailed subdivision of the Jurassic sediments and their correlation. The species diversity of endemic genera (Tethyan and Boreal endemics) was considerably lower, and their life period was limited, not exceeding, as a rule, a stage, i.e., a geological age.

The analysis of taxonomic composition of foraminiferal assemblages from different climatic and bathymetric zones allows me to conclude that changes related to the bathymetric zoning are similar to those that are dependent on the geographic (climatic) position of the assemblages in the Jurassic sedimentary basins. The only exception is the foraminiferal community of the tropical littoral and sublittoral zones, which is unique, as compared to assemblages from other bathymetric and climatic zones of the Jurassic sea basins.

ACKNOWLEDGMENTS

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Paleogeography of the Berriasian–Barremian Ages of the Early Cretaceous

M. A. Zharkov*, I. O. Murdmaa, and N. I. Filatova*****

** Geological Institute, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia*
*** Shirshov Institute of Oceanology, Russian Academy of Sciences, ul. Krasikova 23, Moscow, 117218 Russia*
**** Institute of the Lithosphere, Russian Academy of Sciences, Staromonetnyi per. 22, Moscow, 109180 Russia*

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Abstract—Global litholologic–paleogeographic maps are compiled for the Berriasian, Valanginian, Hauterivian, and Barremian ages of the Early Cretaceous. Main features of paleogeography, sedimentation environments in oceans, regularities in distribution of paleogeographic environments in continental margins, spatial position of arid and humid sedimentation settings in continents, and position of latitudinal climatic belts of the Neocomian time are considered. It is noted that clayey–calcareous hemipelagic and pelagic sedimentation prevailed in the Tethys; carbonate-free clayey–siliceous and calcareous pelagic sedimentation was characteristic of the Pacific, whereas hemipelagic terrigenous sediments were mainly accumulated in the Southern Ocean. Morphological–structural lateral series of deep-water trenches, turbidite fore-arc basins, and volcanic belts of island-arcs and continental margins are distinguished at the convergent boundaries between continental and oceanic plates. Five latitudinal climatic belts of the Neocomian time corresponded to the northern circumpolar humid zone with coal deposits, the northern midlatitudinal humid zone with coal–bauxite–kaolinite deposits, the intersubtropical arid zone with evaporites, the southern midlatitudinal humid zone with coal–kaolinite deposits, and the southern humid zone with coal-bearing sequences.

Key words: *paleogeography, Berriasian, Valanginian, Hauterivian, Barremian, hemipelagic and pelagic oceanic environments, continental margins, arid and humid climates, evaporite and coal-bearing basins, climatic belts.*

INTRODUCTION

This paper is the next in a series of publications devoted to characteristics of global lithologic–paleogeographic maps for subsequent Cretaceous ages compiled with the purpose to elucidate the evolution of paleogeographic and paleoclimatic peculiarities in the spatial distribution of sedimentation settings and volcanic belts during the epoch of the warm biosphere. This paper presents lithologic–paleogeographic maps for the Berriasian, Valanginian, Hauterivian, and Barremian ages of the Early Cretaceous (Figs. 1–4). They were compiled using the same method that was applied for previously published maps for the Middle Cretaceous (Zharkov *et al.*, 1995). By means of independent systems of symbols, arid and humid zones in continents, shelves, island arcs, and in the peripheral and central areas of the ocean are distinguished in these maps. The maps are sufficiently informative. They allow us to elucidate peculiarities of the spatial distribution of carbonate, terrigenous, glauconite-bearing, phosphorite-bearing, and chalk sedimentation environments in shelf and epicontinental seas; to demonstrate features of evaporite sedimentation and coal accumulation in continents; to derict patterns of turbidite and black-shale deposition; and to reconstruct the general distribution of hemipelagic and pelagic sediments in oceans. Simultaneously, they provide sufficiently complete information on the distribution of magmatic and sedimentary–volcanogenic rock complexes in the active continental margins, intracontinental areas, and oceans.

It is impossible to consider in one paper all peculiarities of paleogeographic and paleoclimatic confinement of the distinguished sedimentation and volcanism settings. Keeping this in mind, we paid the most attention to the following: (1) to principal paleogeographic features that existed during the first half of the Early Cretaceous and to paleoenvironments of oceanic sedimentation; (2) to the main regularities in distribution of paleogeographic settings in continental margins; and (3) to the spatial distribution of arid and humid sedimentogenesis in continents and to reconstruction of latitudinal climatic belts of the Neocomian time. The reference list includes only those publications that are essential for this paper. All other works that were used in compiling maps were listed in our previous article (Zharkov *et al.*, 1995).

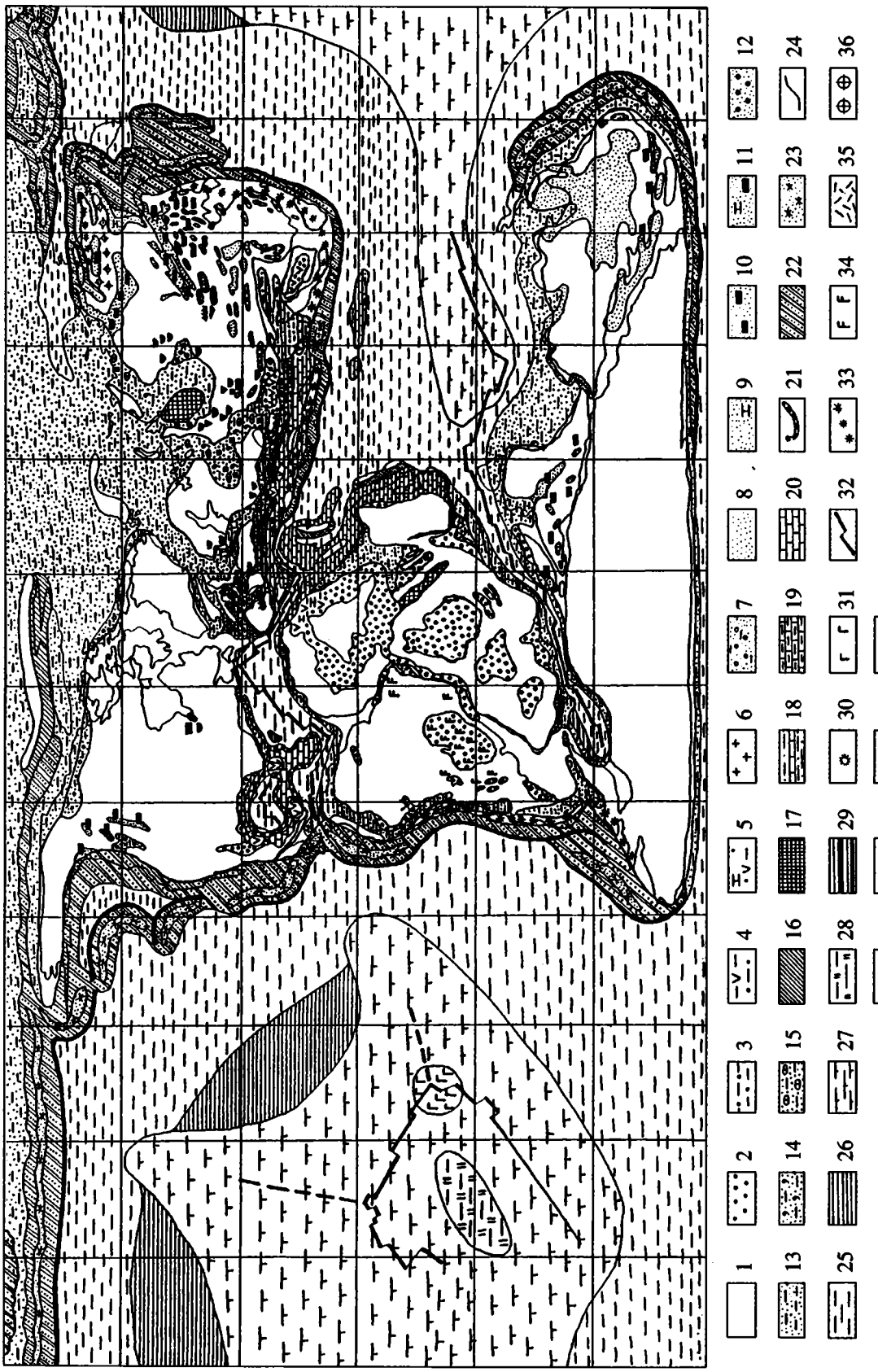
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Fig. 1.

THE PRINCIPAL PALEOGEOGRAPHIC FEATURES AND SEDIMENTATION ENVIRONMENTS IN OCEANS

Paleogeographic peculiarities of the first half of the Early Cretaceous were controlled by the existence of three major continental blocks—Laurasia, Western Gondwana, and Eastern Gondwana, whose position was relatively stable throughout almost the entire Neocomian (Ziegler *et al.*, 1982; Zonenshain *et al.*, 1984; Barron, 1987; Scotese *et al.*, 1987, 1988; Funnell, 1990; Dercourt *et al.*, 1993). Laurasia was located in the northern hemisphere extending from 20°–30° N to the northern polar areas. Western Gondwana, comprising South America and Africa, was mainly situated within the subtropical–tropical belt between 25°–28° N and 35°–40° S. Eastern Gondwana, including Australia, Antarctica, and India, was located in the southern hemisphere south of 35°–40° S and occupied large areas in the southern polar region. In the Berriasian, Eastern Gondwana represented a single continent. In the Valanginian, India started to separate from Australia and later from Antarctica to finally become an isolated continental block in the Barremian (Veevers, 1984; Scotese *et al.*, 1987, 1988; Patriat and Segoufin, 1988). These major continental masses composed the continental hemisphere of the Earth, which opposed the oceanic hemisphere occupied by the Pacific.

Paleogeography of oceanic basins was controlled in the Neocomian by geodynamic processes initiated in the Middle–Late Jurassic. The Tethys, which extended over almost 20 000 km between Laurasia and Gondwana continents from the Caribbean region in the west to the southeast Asian margin and the Australia–Guinea region of Eastern Gondwana in the east continued to open and advance westward. It comprised the eastern Tethys encompassing the area between the Asian part of Laurasia in the north and India and Australia in the south; central (or Mediterranean) Tethys; and western Tethys; where the Central Atlantic (Atlantic Tethys), Gulf of Mexico, and Caribbean regions can be distinguished (Dercourt *et al.*, 1993). At the beginning of the Early Cretaceous, the seaway between North and South America appeared and resulted in formation of the global western current in the Tethyan and Pacific tropical latitudes of the northern hemisphere (Luyendyk *et al.*, 1972; Berggren and Hollister, 1974).

The Mediterranean and western Tethys represented a system of deep basins fringed by shallow pericratonic carbonate platforms and terrigenous shelves; the basins were also separated in some areas by groups of relatively small isolated carbonate platforms (Dercourt *et al.*, 1985; Dercourt *et al.*, 1993; Bogdanov *et al.*, 1994). Pericratonic and isolated carbonate platforms were areas of intense accumulation of biogenic carbonates; from the outer side, they were usually bounded by reefal carbonate buildups (Wilson, 1975; Masse and Philip, 1981; Wilson *et al.*, 1984; Frazier and Schwimmer, 1987; Schlager and Philip, 1990; Masse, 1992). Such carbonate platforms are often referred to the Urgonian type (Masse, 1992). Pericratonic carbonate platforms extended along the eastern margin of Africa, eastern and northern Arabia, and the northern and northwestern coast of Africa. In the southern periphery of Laurasia, they are traceable along the North American margin, in the southern and southwestern coast of the Gulf of Mexico near Baltimore Canyon and in the Blake Plateau, and also in the northern Tethyan margin around Iberia and from the Pyrenees platform to the Carpathian–Balkan region and further eastward to Elbrus and Central Afghanistan (Viniegra, 1981; Seibold, 1982; Young, 1983; Wilson *et al.*, 1984; Dercourt *et al.*, 1985; Tucholke and McCoy, 1986; Zonenshain *et al.*, 1987; Schwimmer, 1987; Schlee *et al.*, 1988; Rakus *et al.*, 1988, 1989, 1990; Watts and Blome, 1990; Masse, 1992; Dercourt *et al.*, 1993). Isolated carbonate platforms were mainly concentrated in two areas, in the Caribbean basin of the western Tethys and in the Mediterranean Tethys. Particularly remarkable were compact groups of isolated carbonate platforms in the Bahamas zone and near the eastern margin of the Med-

Fig. 1. The lithologic–paleogeographic map for the Berriasian age of the Cretaceous.

(1) land; (2–6) deposits of alluvial–proluvial plains, intermontane depressions, lakes, sebkhas, and lagoons in arid zones: (2) red-bed conglomerates, gritstones, and sandstones; (3) red-bed and variegated sandstones, siltstones, and clays; (4) sandstones, siltstones, and clays with gypsum; (5) gypsiferous calcareous, and terrigenous–calcareous sediments; (6) salt-bearing sediments; (7–12) deposits of alluvial and lacustrine–palustrine plains, intermontane depressions, coastal plains intermittently flooded by the sea, and lagoons: (7) gray conglomerates, gritstones, and sandstones; (8) gray sandstones, siltstones, and clays; (9) calcareous–terrigenous sediments; (10) terrigenous coal-bearing (intracontinental) sequences; (11) calcareous–terrigenous coal-bearing (maritime) sequences; (12) terrigenous red-bed carbonate-free sediments; (13–21) deposits of shelf and epicontinental seas: (13) sandstones, siltstones, and clays; (14) glauconite-bearing sediments; (15) phosphorite-bearing sediments; (16) turbidites of shelf slopes and back-arc basins; (17) carbonaceous clayey, clayey–calcareous, and calcareous–siliceous sediments (black shales); (18) siltstones, clays, and limestones; (19) clayey limestones and marls; (20) carbonate platforms; (21) reefs; (22–25) deposits of the continental and island-arc slope and peripheral zones of oceans: (22) turbidites; (23) calc-alkaline and tholeitic basalts and also island-arc terrigenous–volcanogenic rock complexes; (24) deep-sea trenches; (25) hemipelagic clayey, calcareous–clayey, calcareous sediments; (26) pelagic (red) clays; (27) pelagic calcareous and siliceous–calcareous sediments; (28) pelagic siliceous sediments; (29) carbonaceous clayey, calcareous–clayey, calcareous–siliceous, and siliceous sediments (black shales); (30) carbonate atolls; (31) intraplate alkalic and tholeitic basalts; (32) midocean ridges with tholeitic lavas; (33) calc-alkaline magmatic rocks of volcanic–plutonic associations in continental margins; (34) intraplate alkalic, tholeitic, and bimodal volcanic series of continents; (35) distal ash belts; (36) belts of S-granitoid massifs in the collision sutures; (37) bauxites; (38) kaolinite clays, kaolin-bearing rocks; (39) iron ores; (40) boundaries between lithologic complexes and paleogeographic regions.

Fig. 2. The lithologic–paleogeographic map for the Valanginian age of the Cretaceous (symbols as in Fig. 1).

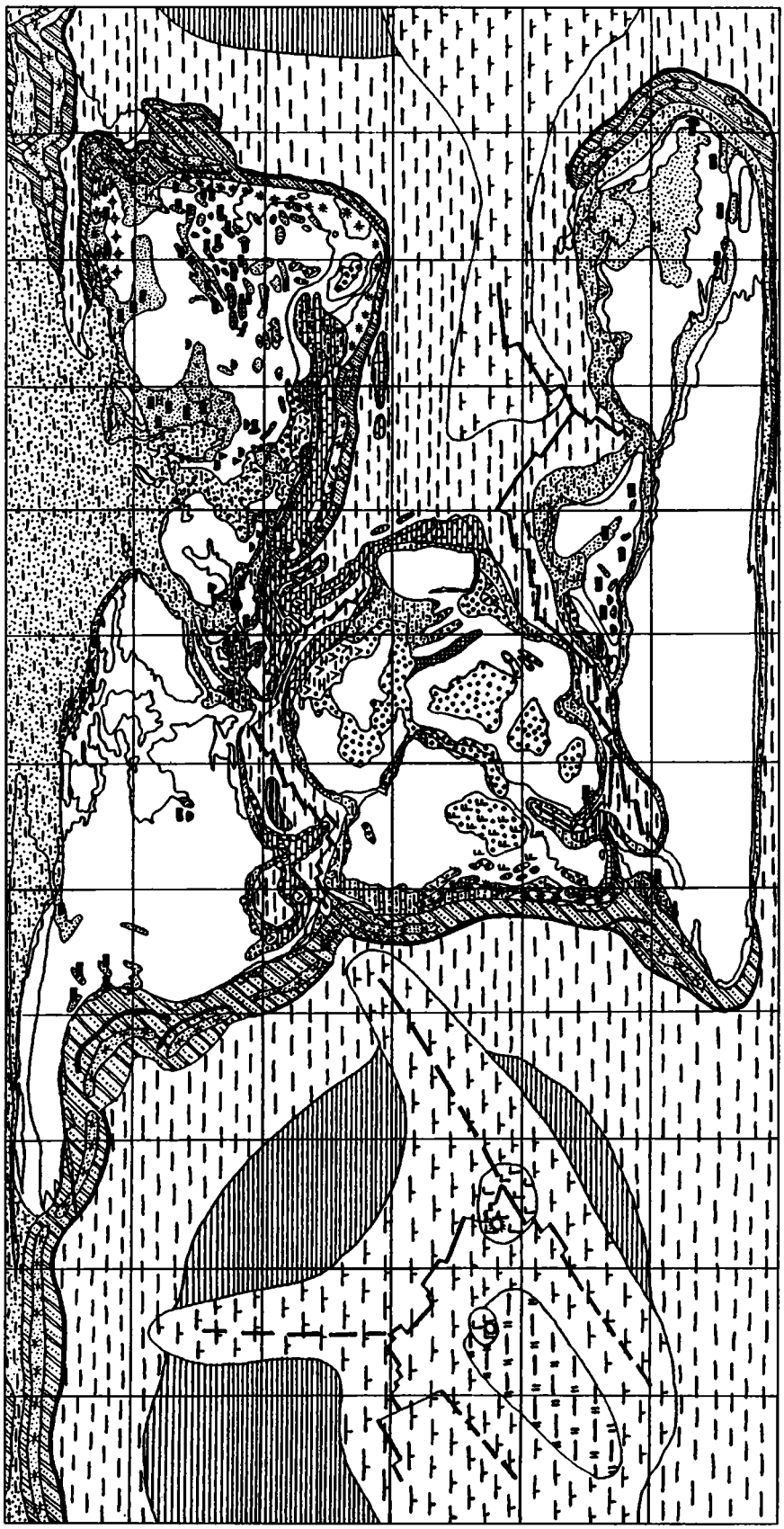

Fig. 3. The lithologic–paleogeographic map for the Hauterivian age of the Cretaceous (symbols as in Fig. 1).

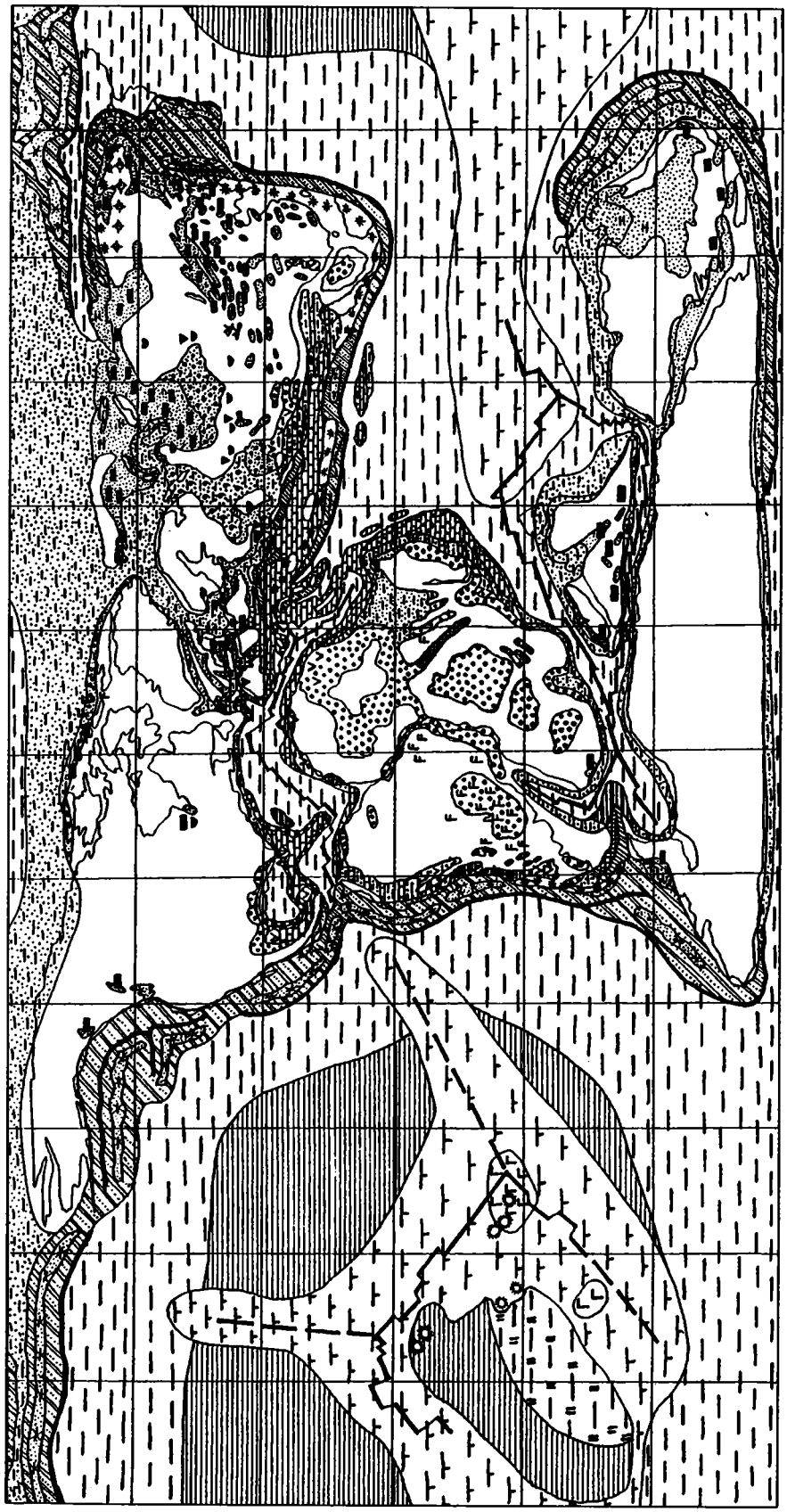

Fig. 4. The lithologic–paleogeographic map for the Barremian age of the Cretaceous (symbols as in Fig. 1).

iterranean Tethys, where they crossed almost the entire basin in the meridional direction and thus created a barrier to westerly surface currents in the tropical belt of the northern hemisphere. Isolated carbonate platforms (Apennines, Apulian, Gavrovo, Taurus, Midian, Bahamas, Mayan, and others) were separated from each other and from pericratonic carbonate platforms by narrow and deep basins, which accumulated either marly–calcareous sediments enriched in planktonogenic organic matter and enclosing the pelagic ammonite fauna or the hemipelagic and pelagic limestones of the Majolica facies and turbidites (Bernoulli, 1972; Bourbon, 1978; Emery and Uchupi, 1984; Dercourt *et al.*, 1985; Dercourt *et al.*, 1993). Because of their small dimensions, some pericratonic and isolated carbonate platforms and the deep basin separating them are not shown in lithologic–paleogeographic maps. All these areas are arbitrarily defined as settings of the shallow-water shelf calcareous sedimentation demonstrating the principal location of the carbonate platform areas.

In the Central Atlantic, Gulf of Mexico, and Caribbean areas, deep-water sedimentation prevailed throughout the Berriasian–Barremian time of the Early Cretaceous. Gray thin-bedded micritic limestones and marls, calcareous hemipelagic sediments, and pelagic limestones were accumulated in these areas (Lancelot *et al.*, 1978; Murdmaa *et al.*, 1979; Tucholke *et al.*, 1979; Owens, 1983; Emery and Uchupi, 1984; Tucholke and McCoy, 1986; Schlee *et al.*, 1988; Stephan *et al.*, 1990; Salvador, 1991; Dercourt *et al.*, 1993). Judging from the difference between modern hypsometric positions of deep-water facies and coeval shallow-water limestones (Blake Plateau), these basins were up to 3–3.5 km deep (Sclater *et al.*, 1977; Murdmaa *et al.*, 1979; Emery and Uchupi, 1984). Clayey–calcareous sediments enriched in organic matter, the deposition of which was most intense during the Valanginian and Hauterivian (Weissert, 1981; Cotillon and Rio, 1984; Emery and Uchupi, 1984; Arthur and Dean, 1986), were characteristic of some areas in the northwestern Atlantic Tethys, Gulf of Mexico, and several limited areas in the Mediterranean Tethys. One of such areas (the northwestern Atlantic Tethys) is arbitrary defined in maps as a basin of black-shale sedimentation.

It was presumed that deep-water hemipelagic facies of clayey–calcareous sediments were also widespread in the eastern part of the Tethys (Dercourt *et al.*, 1993), though geological records about this spacious oceanic basin were almost wholly destroyed by subsequent subduction and collision between Hindustan and Asia. The hypothetical belt of pelagic calcareous sediments is also shown along the reconstructed spreading ridge in the central part of the eastern Tethys.

During the Berriasian–Barremian, the gradual opening of the Southern Ocean and sea-floor spreading advancing southwestward separated Gondwana into two major western and eastern continental blocks. At the initiatial opening stage, the South Atlantic represented a narrow gulf of the Southern Ocean (Rabinowitz and La Brecque, 1979; Scotese *et al.*, 1987, 1988; Patriat and Segoufin, 1988). Mainly terrigenous and calcareous–terrigenous hemipelagic sediments accumulated in the narrow though relatively deep Southern Ocean. In the Hauterivian and Barremian, black shales enriched in organic matter were deposited along its western margin (Malumian *et al.*, 1983; Thomson, 1983; Krasheninnikov and Basov, 1985; Kavun and Vinnikovskaya, 1993).

Beginning in the Barremian, Hindustan started to move away from Antarctica, because a spreading zone appeared between them and created the deep-water seaway. This event marked the initial opening of the Indian Ocean, and its floor became a new area of hemipelagic terrigenous deposition (Patriat and Segoufin, 1988).

In the Neocomian, the Pacific represented a spacious and deep oceanic basin exceeding in size its modern area. According to reconstructions (Zonenshain *et al.*, 1984; Kononov, 1989) and available new data, four major lithospheric plates (Kula, Farallon, Phoenix, and Pacific) continued to evolve here along the system of spreading ridges with two triple-junctions. Reliable data on sedimentation environments are obtained (from deep-sea drilling) only for the Pacific plate, whereas three other plates are completely eliminated by subduction, and their history can be depicted only from isolated blocks (terranes) incorporated into accretionary structures of surrounding continents. The sediment distribution in these vanished plates is shown hypothetically by analogy with the Pacific plate and on the basis of general regularities of oceanic sedimentation.

During the Neocomian, the small Pacific plate born in the Jurassic was located in the southern hemisphere, in the center of the oceanic pelagic area far from the continents, and this resulted in accumulation of typical pelagic facies, whose characteristic features are red coloration, prevalence of biogenic and authigenic components, absence of terrigenous material (except for the finest clay), insignificant thicknesses, and a low sedimentation rate (Murdmaa, 1987).

In the central part of the Pacific plate most distant from the spreading axis, an oceanic area below the carbonate compensation depth (CCD) existed from the Berriasian through the Barremian. Red carbonate-free clayey–siliceous (radiolarian) sediments were accumulated in this part of the basin. This area of abyssal pelagic facies gradually widened during the indicated period of time. In the Barremian, typical pelagic clays including zeolite varieties were accumulated together with radiolarites.

Pelagic calcareous (nannofossil–foraminiferal) oozes were accumulated above the CCD along the mid-oceanic ridges and on submarine rises existing from the Jurassic. On the Schatsky Rise, the Berriasian–Valanginian(?) was marked by the accumulation of black siliceous sediments enriched in organic matter.

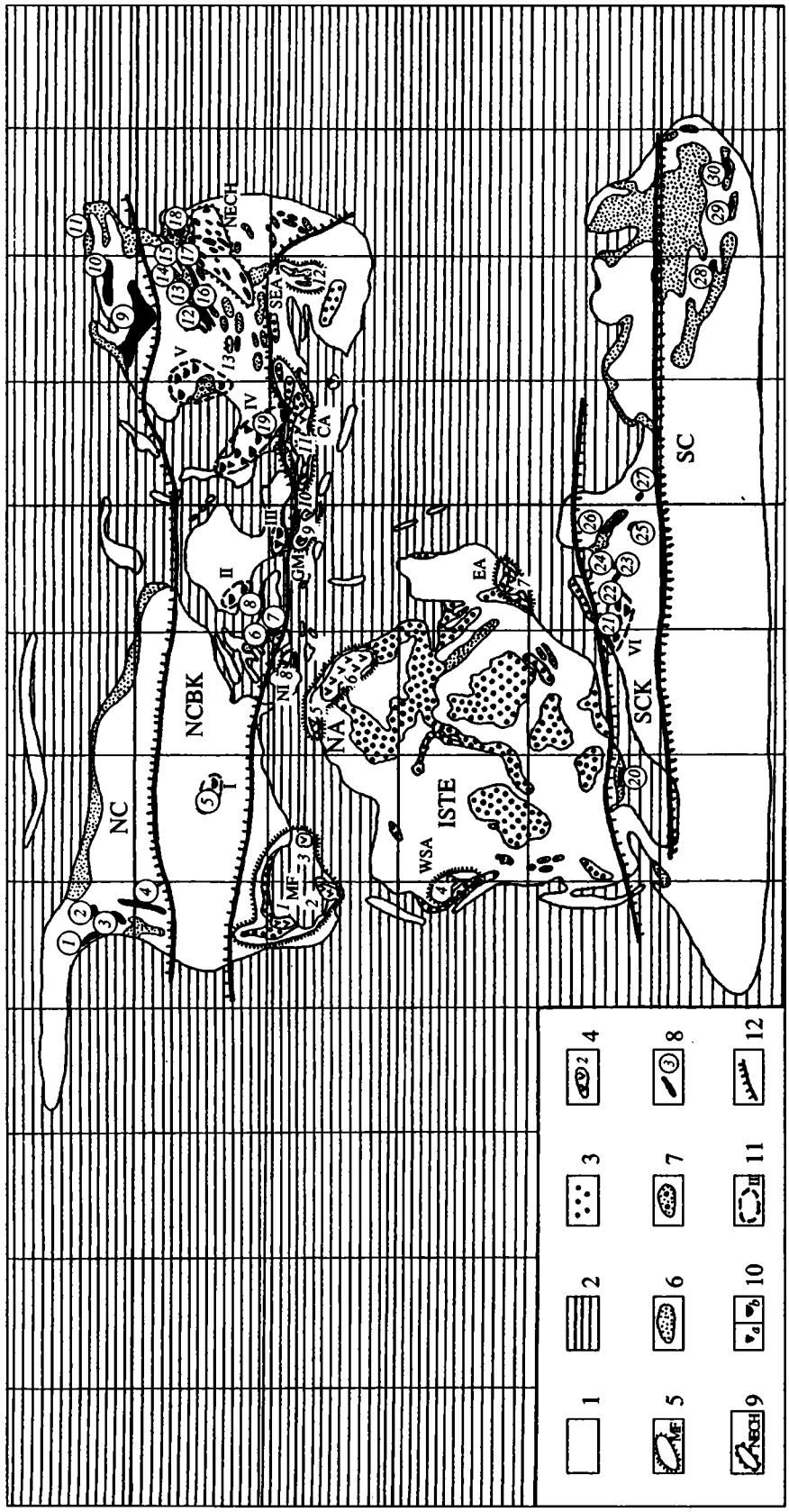
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Fig. 5.

There were several areas of active intraplate volcanism, the largest of which—the Darwin Rise—was situated during the Neocomian near the spreading axis, probably in an area of triple-junction at the eastern corner of the Pacific plate. Volcanic edifices that appeared there in the Hauterivian–Barremian now compose the constituents of the Mid-Pacific Mountains. Volcanoes of the Japanese guyot group in the northern part of the plate, as well as those of the Schatsky Rise, were probably active in earlier epochs as well. As a result of intraplate volcanism, the Ontong-Java and Hess rises were formed in the Barremian. Many volcanic islands rose above the sea level and were subsequently (beginning in the Hauterivian–Barremian) transformed into atolls (Sager *et al.*, 1993).

MAIN DISTRIBUTION PATTERNS OF PALEOGEOGRAPHIC ENVIRONMENTS IN CONTINENTAL MARGINS

This section is dedicated to characteristics of paleogeographic environments in active continental margins of the western, eastern, and southern parts of Laurasia and in marginal areas of Gondwana, except for those adjacent to the Central Atlantic, Tethys, and Southern Ocean (Figs. 1–4).

As was mentioned previously (Zharkov *et al.*, 1995), the method of compiling paleogeographic and paleogeodynamic maps of the ocean–continent transition zone for successive ages of the Early Cretaceous included the analysis of Mesozoic orogenic belts, which fringe continents and are composed of tectonically imbricated nappes and thrust sheets of heterogeneous and heterochronous rock complexes—the terranes of oceanic (sedimentary, ophiolitic, island-arc) and marginal continental formations. The main purpose of this analysis was to determine the primary nature of allochtonous terranes occurring in modern orogenic belts as highly dislocated structural units.

The ocean–continent transition zones considered below encompass peripheral parts of oceanic basins (often comprising chains of volcanic island arcs) and marginal areas of continents. Interaction of oceanic and continental plates along convergent boundaries was responsible for the conformable spatial distribution of paleogeographic settings in the ocean–continent transition zone. As a result, lithologic complexes of different genesis in the peripheral oceanic zones and continental margins show the lenticular–banded distribution in the plan (Figs. 1–4).

In the Berriasian through the Barremian, almost the entire periphery of the paleo-Pacific was fringed by systems of volcanic island arcs, which were connected along the strike with volcanic belts of continental margins in some regions. The lateral succession of structures in the ocean–continent transition zone commonly included a deep-sea trench, a fore-arc basin with turbidite deposits, a volcanic island arc with volcanic rocks of the calc-alkaline and tholeitic series and also with terrigenous–volcanogenic deposits, and a back-arc basin filled with turbidites. Continental margins adjacent to the peripheral oceanic zones were different in their paleogeographic environments. Land areas alternated with shelf seas, which accumulated sandy–clayey and, less commonly, terrigenous–calcareous sediments. In addition, some segments of the circum-Pacific belt were occupied by chains of land volcanoes of marginal continental volcanic chains. These volcanic belts were accompanied by fore-arc depressions filled with turbidite or sandy–clayey sediments.

The small back-arcs occupied by basins were the accumulation areas of red-bed molasses and gray-colored terrigenous and calcareous deposits often in combinations with alkalic basaltic rocks of the intraplate type. In rare cases, such basins were transformed into marginal seas where turbidites and tholeitic basalts that were similar to MORB lavas accumulated.

Characterizing more thoroughly paleogeographic environments in the transition zone from the past

Fig. 5. Belts and provinces of arid and humid sedimentation of the Berriasian Age of the Early Cretaceous.

(1) land; (2) oceans, shelves, and epicontinental seas; (3) provinces and basins of the red-bed arid sedimentation; (4) evaporite basins (*1*, Sabians; *2*, Yucatan; *3*, South Floridan; *4*, Acre; *5*, Moroccan; *6*, Algerian–Tunisian; *7*, Mandera; *8*, Soria; *9*, Moesian; *10*, Georgian; *11*, Central Asian; *12*, Lanpang–Simao; *13*, Dzabhan, Banernur); (5) evaporite provinces and their indices (MF, Mexico–Floridan; WSA, Western South American; NA, North African; EA, East African; NI, North Iberian; GM, Georgian–Moesian; CA, Central Asian; SEA, Southeast Asian); (6) provinces, areas, and basins of terrigenous humid sedimentation (gray beds); (7) humid provinces and areas of variegated and red-bed sedimentation; (8) coal-bearing basins: *1*, Saint Elias; *2*, Whitehorse; *3*, Bowser, Sastus, Skeena; *4*, Foothills and Front Ranges of Rocky Mountains; *5*, Moose River; *6*, Celtic, Bristol, Wild, and Channel; *7*, Parisian; *8*, Western Netherlands, Lower Saxonian, and Altmark–Branderburg; *9*, Lena; *10*, Zyryanka; *11*, Pegtymel'; *12*, West Transbaikalian (Gusinoe Ozero, Uda, Khilok–Chikoi, Eravnoe, and others); *13*, Olekma–Vitim (Ukshum, Vitim, and others); *14*, Southern Yakutia; *15*, Udsk; *16*, East Transbaikalian (Chikoi, Chita–Ingoda, and others); *17*, Amur–Zeya; *18*, Bureya; *19*, Karakamys; *20*, Algoa; *21*, Sakoa; *22*, Palar and others; *23*, Eluri, Ongole, and others; *24*, Wardha, Nagpur, and others; *25*, Talcher; *26*, Narmada (Satpura and others); *27*, Damodor and others; *28*, Otway; *29*, Bass; *30*, Gippsland, Strzelecki; (9) Northeastern Chinese (NECH) coal-bearing province (Songliao, Hailar, Erlian, and other basins); (10) areas of bauxite formation (*a*) and kaolinite-formation (*b*); (11) provinces of kaolinite or kaolinite–bauxite formation: I, Moose River; II, Western Baltic; III, Northern Black Sea; IV, Central Asian–Western Siberian; V, Eastern Siberian; VI, Southern Madagascar–Southern Hindustan; (12) boundary between climatic belts. Climatic belts: NC, northern coal-bearing of the circumpolar humid zone; NCBK, northern coal–bauxite–kaolinite belt of the humid zone in middle latitudes; ISTE, intersubtropical evaporite belt of the arid zone; SCK, southern coal–kaolinite belt of the humid zone in middle latitudes; SC, southern coal-bearing belts of the humid zone.

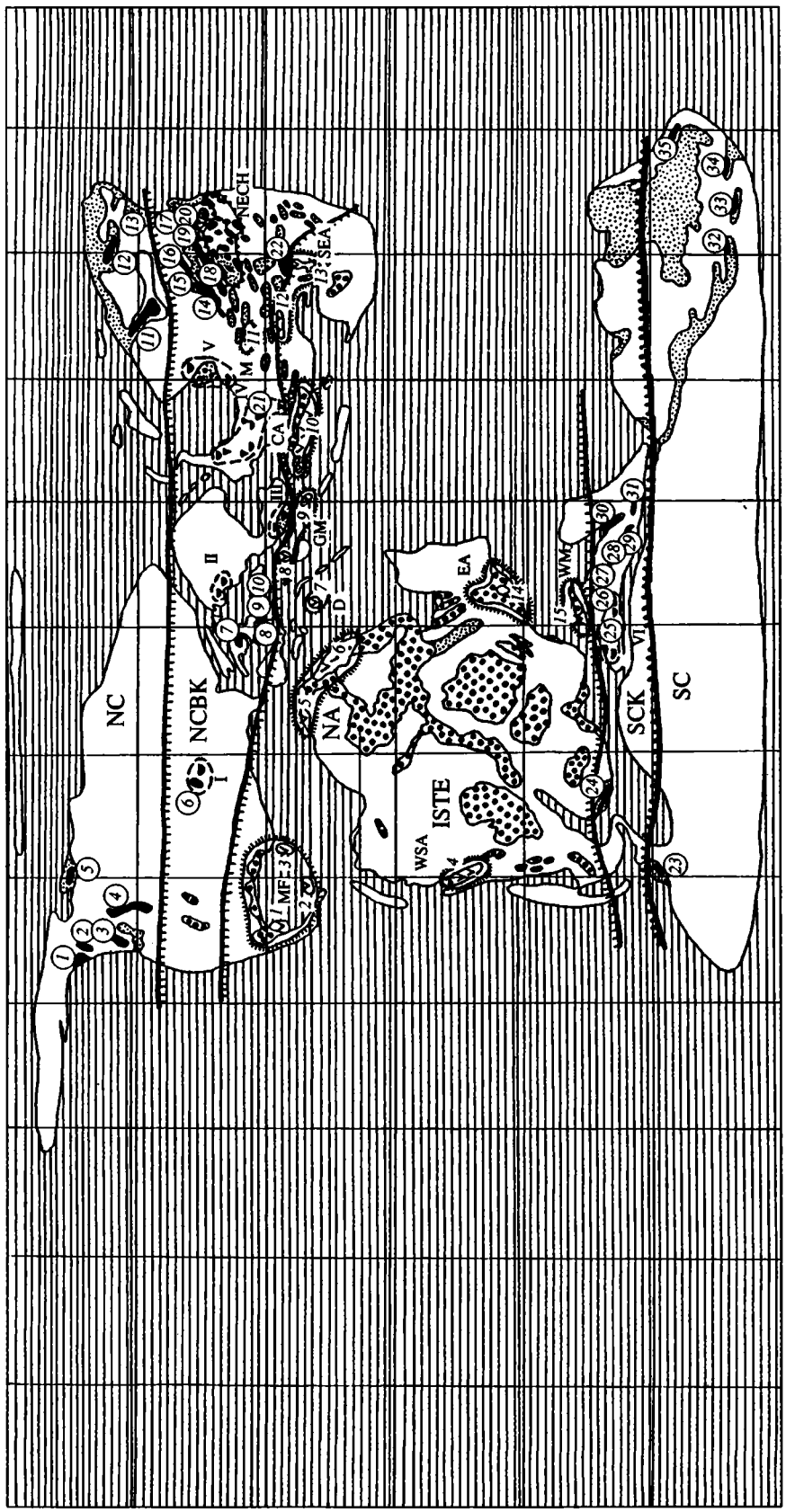
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Fig. 6.

Pacific ocean to its continental margins, we should note that a complex segmented system of volcanic island arcs evolved during the Neocomian along the western coasts of Laurasia (Csejtev *et al.*, 1982; Monger *et al.*, 1982; Shervais and Kimbrough, 1985; Bogdanov and Dobretsov, 1987; Frazier and Schwimmer, 1987; Pindell *et al.*, 1988; Wallace *et al.*, 1989; Miller and Hudson, 1991; Samson *et al.*, 1991; Tardy *et al.*, 1991; Underschulz and Erdmer, 1991; Wilson *et al.*, 1991; Currie and Parrish, 1993). Deep-sea trenches that accompanied these volcanic arcs were locally situated on both the oceanic and the continental sides of island arcs (Wallace *et al.*, 1989). Chains of the Early Cretaceous island arcs along the western coast of Laurasia appeared as early as in the Jurassic time. In the Late Jurassic, some of their segments experienced collision and were integrated into more extended systems (Harper and Wright, 1984; Wallace *et al.*, 1989). In addition to active volcanic island arcs, an underwater chain of the superterrane Talkeetna consisting of dead volcanic arcs of the Paleozoic and Early Mesozoic age existed in the Early Cretaceous along the northeastern periphery of the Mesozoic Pacific Ocean (Jones *et al.*, 1986; Wallace *et al.*, 1989). The Berriasian–Barremian time was marked by the gradual convergence of segments of volcanic island arcs with the Laurasian margin, which continued until the Aptian, when they were amalgamated to the continent (Lanphere *et al.*, 1978; Csejtev *et al.*, 1982; Armstrong *et al.*, 1986; Vaughan, 1995; Zharkov *et al.*, 1995; Filatova, 1996). In the southeastern peripheral part of the ocean, a complex system of island-arc and marginal continental volcanic belts evolved during that time (Figs. 1–4). The northern segment of this system (10°–15° S) included the volcanic island arc situated near the northwestern margin of Gondwana (Colombia, Ecuador, Peru), when volcanoes erupted boninitic and tholeitic lavas (Megard, 1987; Wallrabe-Adams, 1990; Van Thournout *et al.*, 1992). Southward, the Chile–Argentine margin of Gondwana was an area of mountainous landscape. A volcanic belt with thick calc-alkaline volcanic sequences existed there (Coira *et al.*, 1982; Lomize, 1983). A trough with tholeitic and alkalic basalts appeared in the rear of the belt (Dalziel *et al.*, 1974; Suarez, 1979; Grier *et al.*, 1991). This belt continued farther to the south into the Patagonia–Antarctic segment. However, the back-arc depression of this structure represented a marginal sea basin, which accumulated turbidites and MORB-type basalts.

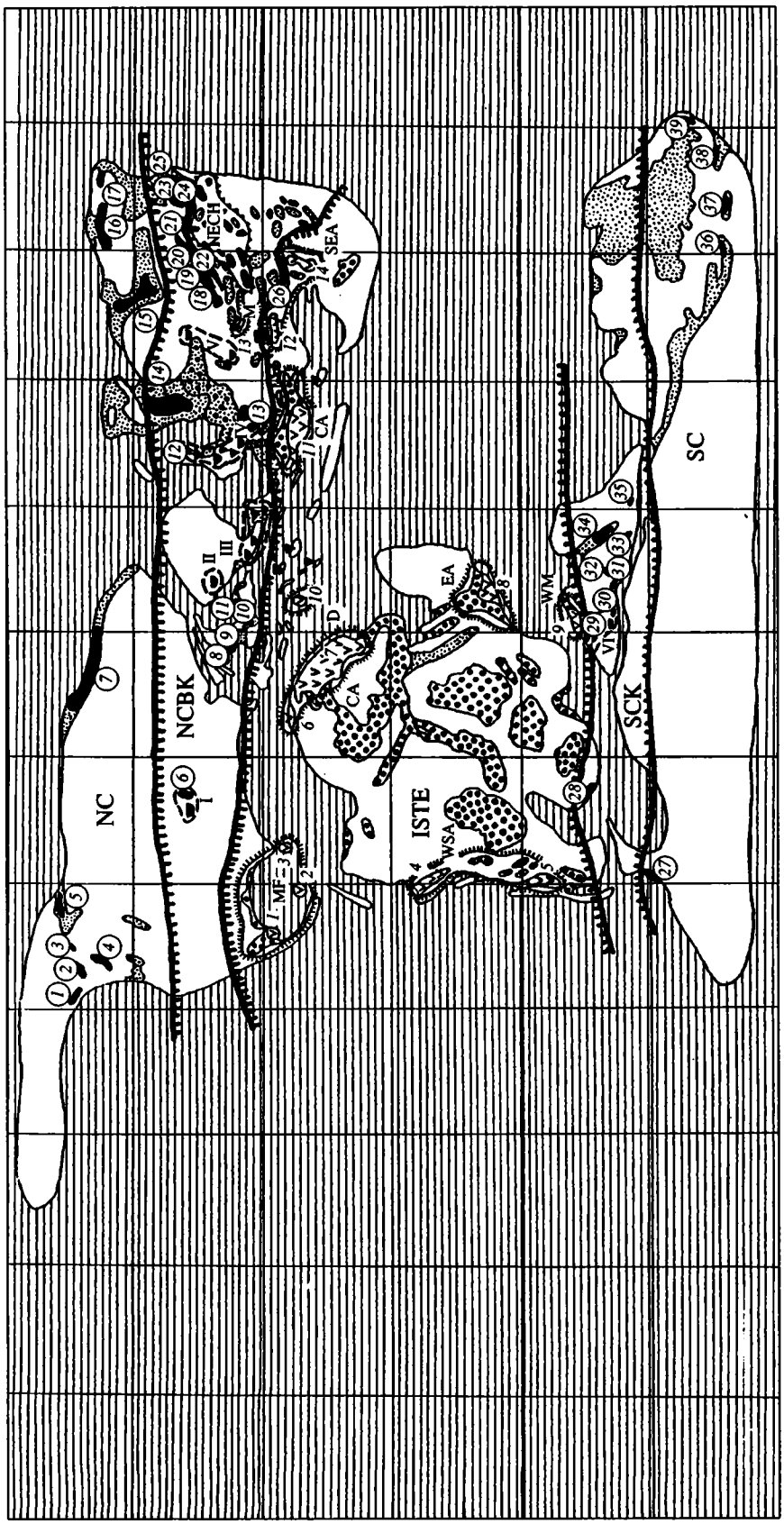
Along the New Zealand–Australia margin of Gondwana in the southwestern periphery of the Mesozoic Pacific ocean, the active volcanic island arc of that time (Lundbrook, 1978; Howell, 1980; Swarko *et al.*, 1983) was conjugated with the deep-sea trench (Figs. 1–4) and surrounded by narrow zones of turbidite sedimentation. The continental margin was occupied here by a shelf sea.

In the northwestern Pacific, the narrow South Anyui sea gulf separating the system of Chukchi continental blocks from the Kolyma–Omolon part of Laurasia existed during the Berriasian–Barremian period. This sea gulf, which gradually narrowed by the mid-Cretaceous time (Filatova, 1988, 1995; Bogdanov and Til'man, 1992; Parfenov *et al.*, 1993), hosted a winding chain of island arcs, which joined the island arcs of the Koryak and Far East regions of the northwestern circum-Pacific margins (Bogdanov, 1988; Faure *et al.*, 1988; Kojima, 1989; Zonenshain *et al.*, 1990; Natal'in and Faure, 1991; Sokolov, 1992).

The Chinese–Korean margin of Laurasia represented in the Neocomian time a land of ragged topography with an extended chain of land volcanoes (Wang and Lin, 1986; Wu and Pei, 1988; Filatova, 1990, 1991). Its rear parts were occupied by depressions with red-bed, evaporite, and, often, coal-bearing deposits. The seaward part of the volcanic belt was fringed by the fore-arc basin and deep-sea trench.

The southern margin of Laurasia in western and eastern Tethys had a similar evolutionary trend during the Neocomian time (Knipper, 1985; Zonenshain *et al.*, 1987; Bogdanov *et al.*, 1994). Spacious areas of this margin were occupied by the epicontinental sea with calcareous sedimentation. The edge of the continent was fringed by the Pont–Transcaucasus ensialic volcanic island arc with turbidites accumulated in the fore-

Fig. 6. Belts and provinces of arid and humid sedimentation in the Valanginian Age of the Early Cretaceous (symbols as in Fig. 5) *Evaporite basins: 1*, Sabinas; *2*, Yucatan; *3*, South Floridan; *4*, Altiplano (Chicamos, Hunin, and others); *5*, Moroccan; *6*, Algerian–Tunisian; *7*, Dinarids; *8*, Moesian; *9*, Georgian; *10*, Central Asian; *11*, Dzabhan; *12*, Weihe; *13*, Lanpang-Simao; *14*, Mandera; *15*, Murundava, Mazhunga. *Evaporite provinces:* MF, Mexico–Floridan; WSA, Western South American; NA, North African; D, Dinarids; GM, Georgian–Moesian; EA, East African; CA, Central Asian; SEA, Southeast Asian; WM, Western Madagascar. *Coal-bearing basins: 1*, Saint Elias; 2, White Horse; *3*, Bowser, Sastus, Skeena, and others; *4*, Foothills and Front Ranges of Rocky Mountains; *5*, Eastern Mackenzie; *6*, Moose River; *7*, Celtic; *8*, Bristol, Wild, Channel; *9*, Parisian; *10*, Western Netherlands, Lower Saxonian, Altmark–Branderburg; *11*, Lena; *12*, Zyryanka; *13*, Northern Okhotsk; *14*, West Transbaikalian; *15*, Olekma–Vitim; *16*, Southern Yakutia; *17*, Udsk; *18*, East Transbaikalian; *19*, Amur–Zeya; *20*, Bureya; *21*, Karakamys; *22*, Weihe; *23*, Alexander; *24*, Algoa; *25*, Sakoa; *26*, Palar and others; *27*, Eluri, Ongole, and others; *28*, Wardha, Nagpur, and others; *29*, Talcher; *30*, Narmada (Satpura and others); *31*, Damodor and others; *32*, Otway; *33*, Bass; *34*, Gippsland, Strzelecki; *35*, Maryborough and Miscellaneous; (NECH) Northeastern Chinese coal-bearing province. *Provinces of kaolinite and kaolinite–bauxite formation*: I, Moose River; II, Western Baltic; III, Northern Black Sea; IV, Central Asian–Western Siberian; V, Eastern Siberian; VI, Southern Madagascar–Southern Hindustan. *Climatic belts*: NC, northern coal-bearing of the circumpolar humid zone; NCBK, northern coal–bauxite–kaolinite belts of the humid zone in middle latitudes; ISTE, intersubtropical evaporite belts of the arid zone; SCK, southern coal–kaolinite belts of the humid zone in middle latitudes; SC, southern coal-bearing belts of the humid zone.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arc basin of its seaside part. The deep-sea trench, whose outer oceanic slope was in contact with hemipelagic and calcareous–clayey deposits of the Tethys, occupied the most frontal position. The incipient Black Sea and South Caspian marginal basins, which appeared as early as in the Late Jurassic, existed in the rear of the Pont–Transcaucasus volcanic arc (Zonenshain *et al.*, 1987; Kaz'min *et al.*, 1987; Dercourt *et al.*, 1993).

Further eastward, in the Tibetan region, the Pontids–Transcaucasus ensialic belt marged into the volcanic island arc of the peripheral zone of the ocean in front of the southern edge of Laurasia, which was bounded by zones of turbidite accumulation (Tapponnier *et al.*, 1981; Bard, 1983; Dietrich *et al.*, 1983; Allegre *et al.*, 1984; Coulon *et al.*, 1986). Even further to the east, the Laurasian margin within the Indochina region of Southeast Asia represented a differentiated mountain land with a chain of volcanoes adjoining the fore-arc trough with turbidite sedimentation and the deep-sea trench (Otsuki, 1985; Wang and Lin, 1986; Poltser and Tapponnier, 1988).

The peculiar feature of landscapes in southern Laurasia and the adjacent peripheral oceanic zone consisted of the presence of undersea rises with carbonate and terrigenous–carbonate sedimentation in the latter zone (Figs. 1–4). These rises are usually considered to represent fragments detached from the African–Arabian plate. The raised blocks (Central Afghanistan, Pamirian, South Tibetan) gradually neared the edge of the Laurasian continent from the Berriasian through the Barremian and collided with the latter in the Aptian–Albian (Allegre *et al.*, 1984).

Thus, the spacious Neocomian ensembles of perioceanic ensialic island arcs existed along the deep-sea trenches of the Mesozoic Pacific, for instance, in its northeastern and northwestern sectors including the southern Anyui gulf. In the southeastern sector of the ocean, the situation was more complicated: the island arc located near northwestern Gondwana merged southward into the central Andean–Antarctic volcanic belt of the continental margin, which further extended along the eastern margin of Gondwana into the southwestern sector of the Pacific. Similarly, island arcs of the northwestern Pacific were replaced southward, in the Chinese–Korean region, by marginal continental volcanic belts.

Volcanic belts along the northern periphery of the Tethys also occupied the marginal position in continents, though segments of oceanic ensimatic island arcs were also formed in some of these areas (for instance, in the North Tibetan region). It is interesting that, in the rear regions of volcanic belts, back-arc and marginal basins existed throughout the Neocomian and tectonic extension in continental margins progressed with time. This can be exemplified by the Rocas-Verdes basin in the Patagonia region of South America, where this period of time resulted in formation of the oceanic-type crust.

The considered paleogeographic settings existed in active margins of Gondwana and Laurasia until 110–105 Ma, i.e., until the cardinal reorganization of lithosheric plates (Knipper, 1985; Bogdanov, 1988; Larson, 1991; Zonenshain and Kuz'min, 1992; Khain and Balukhovskii, 1993; Zharkov *et al.*, 1995; Filatova, 1996).

PECULIAR FEATURES OF SPATIAL DISTRIBUTION OF ARID AND HUMID SEDIMENTATION AREAS

Schemes illustrating the distribution of coal accumulation areas and basins of red-bed or gray-bed sedimentation in the arid and humid zones also show the sites of kaolinite and bauxite formation. They are based on lithologic–paleogeographic reconstructions and also on an analysis and systematization of published data (Bogolepov, 1961; Butov *et al.*, 1962; Douglas, 1964; Martinis and Visintin, 1966; *Atlas litologo–paleogeograficheskikh...*, 1968; Gol'tbert *et al.*, 1968; Busson, 1972; Blant, 1973; Beltrandi and Pyre, 1973; *Poverkhnosti vyravnivaniya...*, 1974; Gevork'yan, 1976; Traves and King, 1976; Kauffman, 1977; Ludbrook, 1978; De Klasz, 1978; Dingle, 1978; Nairn, 1978; Benson *et al.*, 1978; Saint-Mare, 1978; Yasamanov, 1978; *Kory vyvetrivaniya...*, 1979; Megnien, 1980; Monakhov *et al.*, 1981; Parrish *et al.*, 1982; Ziegler, 1982,

Fig. 7. Belts and provinces of arid and humid sedimentation of the Hauterivian Age of the Early Cretaceous (symbols as in Fig. 5). *Evaporite basins*: *1*, Sabinas; *2*, Yucatan; *3*, South Floridan; *4*, Altiplano (Chicamos, Hunin, and others); *5*, Neuken; *6*, Moroccan; *7*, Algerian–Tunisian; *8*, Mandera; *9*, Murundava; *10*, Dinarids; *11*, Central Asian; *12*, Qaidam; *13*, Dzabhan; *14*, Lanpang-Simao. *Evaporite provinces:* MF, Mexico–Floridan; WSA, Western South American; NA, North African; EA, East African; WM, Western Madagascar; D, Dinarids; CA, Central Asian; M, Mongolian; SEA, Southeast Asian; *Coal-bearing basins*: *1*, Saint Elias; *2*, White Horse; *3*, Peel; *4*, Laberge, Bowser, Sastus, and others; *5*, Eastern Mackenzie; *6*, Moose River; *7*, Sverdrup; *8*, Celtic; *9*, Bristol, Wild, Channel; *10*, Parisian; *11*, Western Netherlands, Lower Saxonian, Altmark–Branderburg; *12*, North Uralian; *13*, Karakamys; *14*, Enisei–Taz; *15*, Lena; *16*, Zyryanka; *17*, Anui; *18*, West Transbaikalian; *19*, Olekma–Vitim; *20*, Southern Yakutia; *21*, Amur–Zeya; *22*, East Transbaikalian; *23*, Udsk; *24*, Bureya; *25*, Partizansk, Razdol'naya; *26*, Central Qinling; *27*, Alexander; *28*, Algoa; *29*, Sakoa; *30*, Palar; *31*, Eluri, Ongole; *32*, Wardha, Nagpur; *33*, Talcher; *34*, Narmada (Satpura and others); *35*, Damodor and others; *36*, Otway; *37*, Bass; *38*, Gippsland, Strzelecki; *39*, Clarence, Moreton, and Miscellaneous; (NECH) Northeastern Chinese coal-bearing province. *Provinces of kaolinite and kaolinite–bauxite formation*: I, Moose River; II, Western Baltic; III, Northern Black Sea–Donets; IV, Uralian–Western Siberian; V, Eastern Siberian; VI, Southern Madagascar–Southern Hindustan. *Climatic belts*: NC, northern coal-bearing belt of the circumpolar humid zone; NCBK, northern coal–bauxite–kaolinite belt of the humid zone in middle latitudes; ISTE, intersubtropical evaporite belt of the arid zone; SCK, southern coal–kaolinite belt of the humid zone in middle latitudes; SC, southern coal-bearing belt of the humid zone.

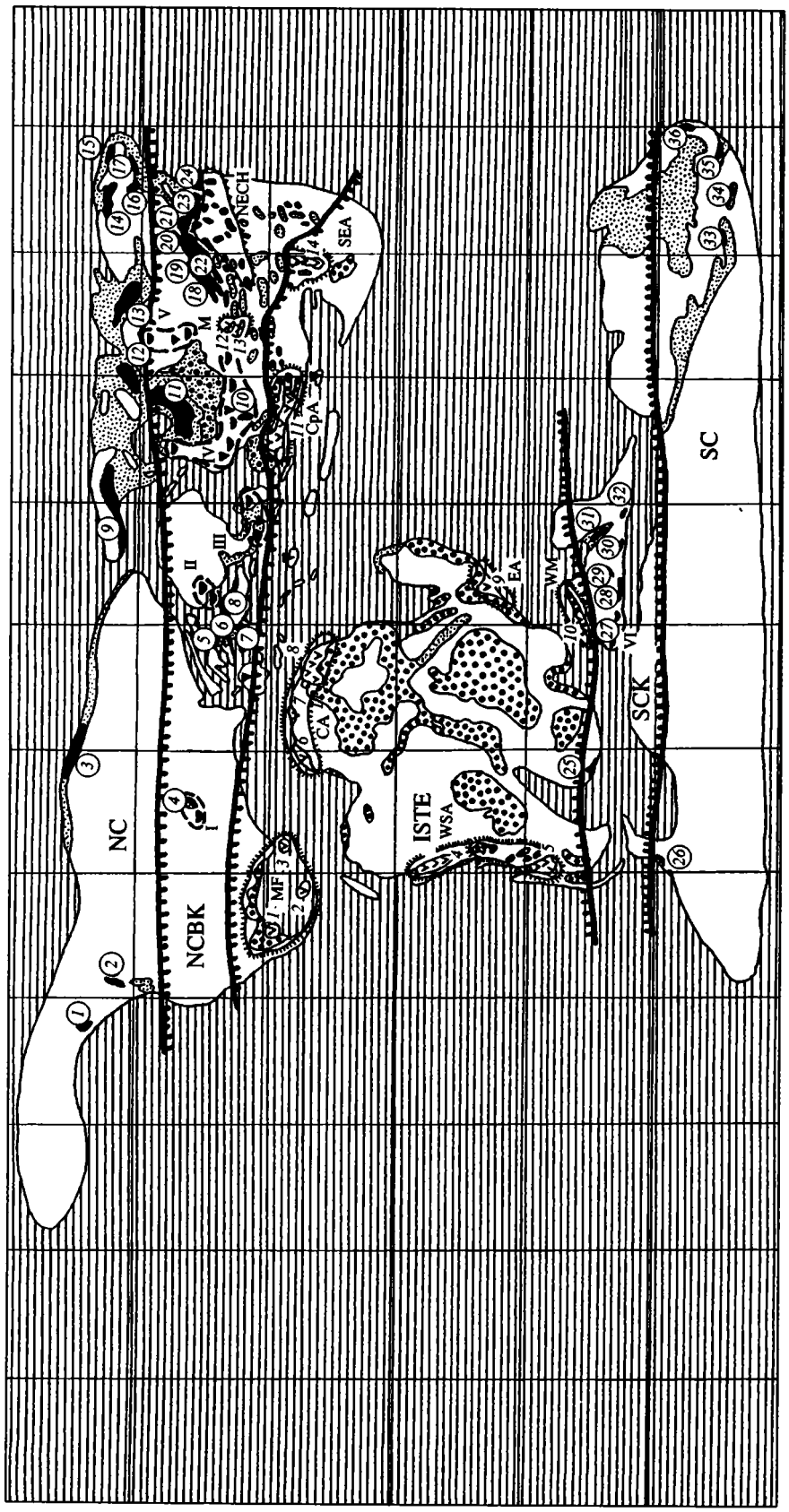

Fig. 8.

1988; Balkwill *et al.*, 1983; Malumian *et al.*, 1983; Dingle *et al.*, 1983; Petri and Mendes, 1983; Thomson, 1983; Emery and Uchupi, 1984; Cahen *et al.*, 1984; *Atlas...*, 1985; Salomon and Drillien, 1985; *Stratigrafiya...*, 1986; Chen Pei-Li, 1987; Frazier and Schwimmer, 1987; Liu Qun *et al.*, 1987; Petri, 1987; Reyment and Dingle, 1987; Riccardi, 1987; Zirsmeister, 1987; Blakey *et al.*, 1988; Butterwarth *et al.*, 1988; Hutchison, 1989; Smith, 1989; Arthur *et al.*, 1990; Bardossy and Aleya, 1990; Bardossy and Dercourt, 1990; Bussert *et al.*, 1990; Luger *et al.*, 1990; Wycisk *et al.*, 1990; Acharyya and Lahari, 1991; Basov and Vishnevskaya, 1991; Wycisk, 1991; *Atlas paleogeograficheskikh...*, 1992; Tashliev and Tovbina, 1992; Dercourt *et al.*, 1993; Kavun and Vinnikovskaya, 1993; Shulgina *et al.*, 1994; Rehakova *et al.*, 1995; Sitian *et al.*, 1995). These schemes allow the main features of zonal arrangement of arid and humid sedimentation environments on continents to be distinguished and the position of climatic belts throughout the indicated time interval to be reconstructed (Figs. 5–8).

In the Berriasian, the environments of arid sedimentation prevailed over all of western Gondwana, along the southern margin of Laurasia, and in adjacent areas of the Tethys (Fig. 5). Various evaporite basins with predominant sulfate sedimentation existed in these regions either as floodplain, lacustrine, limnoalluvial, deltaic, lagoonal, coastal, and continental sebkhas or as evaporite–carbonate platforms. Simultaneously, the red-bed terrigenous sedimentation was characteristic of spacious areas of the West Gondwanan interiors, as well as of the southwestern and southeastern parts of Laurasia.

The extended evaporite-accumulation belt became traceable at that time also along the entire southern margin of Laurasia and northern periphery of the Tethys. It was spread over more than 15 000 km from the Caribbean region in the west to Southeast Asia in the east within the limits of the northern subtropical and tropical zones. The belt comprised a series of isolated evaporite basins located in the Mexican–Floridan, North Iberian, Georgian–Moesian, Central Asian, and Southeast Asian regions, where many of them, e.g., the Yucatan, Sabinas, South Floridan, Soria, Moesian, Georgian, Central Asian, and Lanpan-Simao basins, had variable sedimentation enviroments.

Three isolated regions of Berriasian evaporite sedimentation are distinguished in the territory of western Gondwana. In the western part of the continent, the western province of South America is outlined by the distribution of evaporites of the Akra basin. The spacious North African region of evaporite accumulation encompasses the Moroccan, Algerian–Tunisian, and a series of smaller continental and coastal evaporite basins distinguished in the north. In the east, the extended East African evaporite region is traceable along the past continental margin running through boundary areas of Kenya, Somalia, and Ethiopia and comprising the Mandera and some other smaller basins. Some peculiar features of the spatial distribution of evaporite-accumulation regions in western Gondwana attract attention. For instance, the North African region is located north of the equator, mostly in the northern tropical belt. Two other evaporite regions in western South America and East Africa extend to the southern tropical belt approximately between 10° S and 30° S. It is seen that the North African evaporite region is situated near the South Laurasian evaporite belt, and, in fact, it is a part of the northern tropical to subtropical zone of evaporite sedimentation. In its western half, this zone covered not only the southern areas of Laurasia but also the entire western and central Tethys, along with adjacent areas of Gondwana, whereas in its eastern half, it occupied the southern and southeastern areas of Laurasia and the northern periphery of the eastern Tethys.

These peculiarities in the spatial distribution of the Berriasian evaporite basins and areas reveal their confinement to two latitudinal belts: the northern Tethyan–South Laurasian belt and the southern one, which is tentatively termed as the Southwestern Gondwanan belt encompassing the western South American and East African evaporite-accumulation regions. Environments of continental red-bed sedimentation of the arid type prevailed between these belts and in many interior basins of western Gondwana. They were widespread in many equatorial, tropical, and subtropical areas. Thus, it is evident that almost all of western Gondwana,

Fig. 8. Belts and provinces of arid and humid sedimentation in the Barremian Age of the Early Cretaceous (symbols as in Fig. 5). *Evaporite basins*: *1*, Sabinas; *2*, Yucatan; *3*, South Floridan; *4*, Altiplano (Chicamos, Hunin, and others); *5*, Neuken; *6*, Tindouf–Ayun; *7*, Moroccan; *8*, Algerian–Tunisian; *9*, Mandera; *10*, Murundava; *11*, Central Asian; *12*, Dzabhan; *13*, Banernur; *14*, Lanpang-Simao. *Evaporite provinces*: MF, Mexico–Floridan; WSA, Western South American; NA, North African; EA, East African; WM, Western Madagascar; M, Mongolian; SEA, Southeast Asian. *Coal-bearing basins*: *1*, Saint Elias; *2*, Bowser, Sastus, and others; *3*, Sverdrup; *4*, Moose River; *5*, Celtic; *6*, Bristol, Wild, Channel; *7*, Parisian; *8*, Western Netherlands, Lower Saxonian, Altmark–Branderburg; *9*, Northern Barents; *10*, Karakamys; *11*, West Siberian; *12*, Khatanga; *13*, Lena; *14*, Zyryanka; *15*, Anui; *16*, Omsukchan; *17*, Taigonos; *18*, West Transbaikalian; *19*, Olekma–Vitim; *20*, Southern Yakutia; *21*, Amur–Zeya; *22*, East Transbaikalian; *23*, Bureya; *24*, Partizansk, Razdol'naya; *25*, Algoa; *26*, Alexander; *27*, Sakoa; *28*, Palar, Eluri, Ongole; *29*, Wardha, Nagpur; *30*, Talcher; *31*, Narmada (Satpura and others); *32*, Damodor and others; *33*, Otway; *34*, Bass; *35*, Gippsland, Strzelecki; *36*, Clarence, Moreton, Miscellaneous; (NECH) Northeastern Chinese coal-bearing province. *Provinces of kaolinite and kaolinite–bauxite formation*: I, Moose River; II, Western Baltic; III, Northern Black Sea–Donets; IV, Uralian–Western Siberian; V, Eastern Siberian; VI, Southern Madagascar–Southern Hindustan. *Climatic belts*: NC, northern coal-bearing belt of the circumpolar humid zone; NCBK, northern coal–bauxite–kaolinite belt of the humid zone in middle latitudes; ISTE, intersubtropical evaporite belt of the arid zone; SCK, southern coal–kaolinite belt of the humid zone in middle latitudes; SC, southern coal-bearing belt of the humid zone.

except for its extreme southern areas, was under the influence of the predominantly arid climate of the Berriasian. In general, the Berriasian was the formation period of a single and very wide belt of arid sedimentation, which covered not only subtropical and tropical regions of the northern and southern hemispheres of the Earth but also the equatorial zone of western Gondwana. This arid belt is recognized as the intersubtropical evaporite belt (ISEB). Its northern boundary was near 30° N, and the southern one was approximately between 40° S and 50° S.

The ISEB retained its boundaries almost unchanged throughout the Valanginian, Hauterivian, and Barremian times (Figs. 6–8). Only the number of evaporite basins and areas changed, and the spatial distribution of red-bed arid terrigenous sedimentation insignificantly varied. For instance, beginning in the Valanginian, the evaporite accumulation stopped in the northern Iberian region and, beginning in the Hauterivian, in the Georgian–Moesian region. A new evaporite accumulation area appeared in their place in the central Tethys. Configuration and boundaries of the North African region also somewhat changed, as it became more spacious in the Barremian time owing to the appearance of evaporite sedimentation in the Tindouf–Aayun basin. The configuration of the western South American evaporite region also changed, when the center of evaporite accumulation moved here southward into the Neuken basin during the Hauterivian and Barremian. However, despite all these changes, the spatial distribution of evaporite basins and areas was persistent in its general patterns, and they were steadily confined to two latitudinal belts: the northern Tethyan–South Laurasian and the southern Southwest Gondwanan. During the Valanginian, Hauterivian, and Barremian ages, environments of red-bed arid sedimentation also continued to exist not only in tropical but also in equatorial areas of western Gondwana.

During the first half of the Early Cretaceous, humid areas and belts comprised coal-bearing basins, areas of kaolinite and bauxite formation, and regions of gray-colored terrigenous sediments. The zonal patterns in distribution of areas with humid sedimentation were most clear in the northern hemisphere, within the Laurasian continent. Two latitudinal humid belts were pronounced here through all the Neocomian ages. There were the northern mid-latitude coal–bauxite–kaolinite (NCBK) belt and the northern circumpolar coal-bearing (NC) belts.

The NCBK belt is traced both in the Eurasian and North American parts of Laurasia. In the eastern part of Asia, it corresponds to the spacious East Asian coal-bearing province comprising more than 300 coal-bearing basins formed in a system of parallel and subparallel flow-through freshwater lakes and river valleys and also in internal-drainage depressions separated by extended ridges. The following coal-bearing regions are distinguished within this province: West Transbaikalian (Gusinoe Ozero, Uda, Eravnoe, Hilok-Chikoi, and other basins), East Transbaikalian (the Chikoi, Chita–Ingoda, and numerous smaller coal-bearing basins), Olekma–Vitim (the Ukshum, Vitim, and other basins), Northeastern China (Hailar, Erlian, Songliao, and a number of other coal-bearing basins), and also the South Yakut, Udsk, Amur–Zeya, Bureya, and other regions. The Partizansk and Razdol'naya basins appeared here in the Hauterivian and Barremian, whereas the Weihe and Central Qinling basins near the southern boundary of the belt were formed in the Valanginian and Hauterivian. Thick black-shale bituminous sequences of argillite, siltstone, and marl were intermittently formed during almost the entire Neocomian period in many lake basins (Gusinoe Ozero, Zaza, Eravnoe, Unda–Onon, Songliao, and others). The system of coal-bearing and black-shale lacustrine basins extended far westward to Mongolia.

In the rest of Eurasia, the humid belt of middle latitudes comprised not only the basins of continental and coastal coal-accumulation bùt the provinces and areas of kaolinite- and bauxite-formation as well. Five major kaolinite and kaolinite–bauxite provinces always existed there in the Neocomian, through they changed their configuration and dimensions with time. They were the Moose River, West Baltic, and northern Black Sea region (beginning in the Hauterivian, the Northern Black Sea–Donets province) and also the Central Asian–West Siberian (in the Hauterivian–Barremian, Uralian–West Siberian) and East Siberian provinces. These regions were constantly or, probably, intermittently marked by the formation of lateritic weathering crusts, secondary kaolinites, bauxite-bearing deposits, kaolinite-to-bauxite clays, and kaolinite–quartz sands and also by the accumulation of other reworked weathering products. The Neocomian coal-bearing basins in the central and western areas of Eurasia were mostly confined to the southern zones of the midlatitudinal belt. This can be exemplified by the Celtic, Bristol, Wild, Channel, Paris, Western Netherlands, Lower Saxonian, Altmark–Branderburg, and Karakamys basins. In the Hauterivian–Barremian, coal formation was also characteristic of spacious areas of the West Siberian basin.

In the North America territory, the Neocomian humid belt of northern middle latitudes includes only the Moose River basin, where the Mattagami Formation encloses the Onakawana lignite beds alternating with kaolinite clays and quartz sands.

The humid belt in question was peculiar because of the the simultaneous development of coal-, kaolinite-, and bauxite-formation environments in continents and specific sedimentation in spacious epicontinental seas confined to it (Figs. 1–4). The most characteristic in this respect were the East European, West Siberian, and Enisei–Khatanga seas, which represented typical marine basins of terrigenous sedimentation with autigenic glauconite, phosphate, and iron minerals com-

bined, sometimes with black-shale accumulation (the West Siberian basin of the Berriasian and Valanginian ages).

The southern boundary of the NCBK belt remained almost stable during the entire Neocomian and is now reliably traceable in the central areas of Laurasia. There, it passes near 28°–30° N between closely spaced evaporite provinces on the one side, and coal-bearing, kaolinite, and kaolinite–bauxite provinces on the other side. In eastern Asia, this boundary turns to the south, where the belt comprises a number of continental basins of eastern China with volcanogenic and gray-colored terrigenous sediments of the humid zone. In North America, the southern boundary of the belt is not far north of the Mexican–Floridan evaporite province. The northern boundary of the belt is arbitrarily drawn near 57°–60° N with due regard for the position of both the areas of coal-accumulation and kaolinite-formation and the epicontinental seas with glauconite, phosphate, and iron-oxide terrigenous sedimentation.

The northern circumpolar coal-bearing belt of the Berriasian, Valanginian, Hauterivian, and Barremian ages encompassed marginal regions of Laurasia located north of 57°–60° N. It is established on the basis of distribution of coal-bearing basins there. In the Berriasian–Valanginian, these basins were concentrated in two regions of the opposite continental margins of Laurasia corresponding to northwestern North America and northeastern Asia. The first of these regions comprises the Saint Elias, White Horse, Bowser, Sastut, Skeena, and Eastern Mackenzie basins and coal deposits of the foothills and Front Ranges of the Rocky Mountains. The other belt includes the Lena, Zyryanka, and Pegtymel' basins. In the Hauterivian and, particularly, Barremian, environments of coastal coal accumulation were characteristic of almost the entire territory of coastal regions of the circumpolar belt, where the Sverdrup, North Barents, Khatanga, Anyui, and Omsukchan coal-bearing basins were formed in addition to the basins mentioned above.

The data available for the southern hemisphere are also sufficient for recognizing and tracing two Neocomian humid belts similar to those outlined in the northern hemisphere. The southern coal–kaolinite (SCK) humid belt can be detected in the middle latitudes of the southern hemisphere. It is recognized as comprising many coal-bearing basins, such as the Sakoa, Palar, Eluri, Ongole, Vardha, Nagpur, Talcher, Narmada, and Damodor deposits in the northern regions of eastern Gondwana (southern margins of Madagascar and southeastern and eastern parts of Hindustan), and also the kaolinite and kaolinite-bearing deposits in the spacious territory tentatively termed as the Southern Madagascar–Southern Hindustan province. The southern margin of Africa, where the Algoa coal-bearing basin was discovered, also belongs to this coal–kaoilinite belt. The northern boundary of the belt is quite unambiguously traced along the eastern margin of the Western Madagascar evaporite province, as well as along the periphery of the arid areas with the red-bed deposits in Africa and South America. This boundary is located near 40°–50° S. The southern boundary of the belt is less obvious and arbitrarily placed at approximately 60° S, because only a few areas and basins of coal formation have been located south of this latitude for all the Neocomian ages.

The southern coal-bearing (SC) humid belt is distinguished by the presence of continental and coastal basins of red-bed humid sedimentation in Australia and some areas of the Antarctic continent, also including isolated coal-bearing basins spaced from each other. In the western margin of eastern Gondwana, only one coal-bearing basin, the Alexander, is known in the island of the same name near the West Antarctic coast, whereas a series of the Otway, Bass, Gippsland, Strzelecki, and other coal-bearing basins in Australia is located in the eastern part of Gondwana. Despite the limited number of coal-bearing basins, the peculiarities of their spatial distribution allow us to define rather reliably the southern coal-bearing humid belt with the northern boundary located near 60° N.

CONCLUSION

The presented analysis of original lithologic–paleogeographic maps compiled for the Berriasian, Valanginian, Hauterivian, and Barremian ages of the Early Cretaceous allows the following conclusions to be drawn.

(1) The Neocomian was the period of the final formation of the Tethys Ocean, the sublatitudinal seaway between Laurasia and Gondwana for the global western current. Hemipelagic clayey–calcareous sediments prevailed in the deep Tethyan basins, where shallow-water carbonate platforms and reefs also existed. In the incipient Southern Ocean and also in the narrow seaway between Hindustan and Antarctica that emerged beginning in the Hauterivian, hemipelagic terrigenous sediments were deposited. The central part of the Pacific was separated by a system of midocean ridges (spreading zones) into several deep basins with prevailing red-bed pelagic clayey–siliceous sedimentation, while ridges and submarine rises accumulated pelagic calcareous sediments. Subalkalic basaltic volcanism occurred in several areas of the Pacific plate.

(2) From the Berriasian through the Barremian time, the convergent boundaries of lithospheric plates were represented by a combination of marginal continental and island-arc (ensialic and ensimatic) volcanic belts, in the frontal part of which fore-arc basins and deep-sea trenches with turbidite sedimentation represented the successive lateral morphostructures related to volcanic belts. One of the peculiar environmental features of this period was the moderate crustal extension behind the volcanic belts of continental margins, where it resulted in the formation of back-arc and marginal sea basins.

These environments existed until the Aptian–Albian time, when the Middle Cretaceous orogeny substantially changed the paleogeography of the ocean–continent transition zones.

(3) The revealed spatial distribution of arid and humid sedimentation areas allows us to distinguish five latitudinal climatic belts of the Berriasian–Barremian time: the northern circumpolar coal-bearing humid zone, the northern coal–bauxite–kaolinite belt in the humid middle latitudes, the intersubtropical evaporite belt of the arid zone, the coal-bearing–kaolinite belt of the southern middle latitudes, and the southern coal-bearing belt. Two peculiar features of the Neocomian climatic zoning are interesting. The first, the asymmetric distribution of the Earth's humid belts, is evident. In the northern hemisphere, the humid belt occupies the largest territory of all the regions of Laurasia located north of 30° N, whereas humid areas of the southern hemisphere are displaced to the south beyond 40°–50° S. This was probably related to the particular position of Laurasia and eastern Gondwana in the northern and southern hemispheres, respectively, and also the position of the Tethys in the northern tropical belt. In addition, a single very wide intersubtropical arid belt between 30° N and 40°–50° S is unambiguously defined as the zone of evaporite and red-bed sedimentation in both the tropical and equatorial areas. The prevalence of arid environments in this spacious zone probably resulted from the fact that the vast western Gondwana continent was located near the equator.

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Vegetation and Lake Levels in Northern Mongolia in the Last 12 500 Years as Indicated by Data of Pollen and Diatom Analyses

N. I. Dorofeyuk* and P. E. Tarasov**

**Severtsov Institute of Ecological Problems and Evolution, Russian Academy of Sciences, Leninskii pr. 33, Moscow, 117071 Russia*
***Moscow State University, Moscow, 11899 Russia*

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Abstract—Bottom deposits of the Dood Nuur (51°20′ N, 99°23′ E), Hovsgol Nuur (50°32′ N, 101°10′ E), and Gun Nuur (50°15′ N, 106°36′ E) lakes of northern Mongolia are analyzed and discussed. Data of pollen and diatom analyses are used to reconstruct changes in terrestrial vegetation, the chronological succession of which throughout the last 12 500 years is inferred on the basis of the radiocarbon dating. Taiga forests consisting of spruce, larch, pine, and birch species were widespread in the Darhat basin 12 300–12 500 years ago. A degradation of arboreal vegetation and growing abundance of steppe coenoses took place 10 600–12 200 years ago. The larch–birch–pine forests became dominant again in the Darhat and Hovsgol basins during the Holocene. In the Orhon–Selenga basin, forest coenoses considerably decreased in area 8700–9600 years ago. The pine–larch forests spread over the region to the maximum during the last 8000 years. The lake levels were variable, and these changes did not coincide in the mountain (the Hovsgol and Dood Nuur lakes) and plain regions (the Gun Nuur Lake). The lake level in the first region was low in the middle Holocene (5400–6500 years ago), when the Gun Nuur Lake was at the transgression stage.

Key words: *lake, bottom deposits, pollen and diatom analyses, radiocarbon dating, lake-level fluctuations, late Pleistocene, Holocene.*

INTRODUCTION

Data on environmental changes in the Asian interiors during the late post-glacial period are scarce at present. However, they are of a high importance for specialists in paleogeography and paleoecology who are interested in the modeling and reconstruction of the past climate and vegetation. We chose northern Mongolia as the test area to be studied because of the peculiar environmental situation in this region that includes spacious depressions surrounded by mountain ridges and is located far inside the continent, beyond the boundary of possible influence of the Atlantic and Pacific atmospheric masses. We believe that the variable (from higher to lower) activity of moisture-bearing atmospheric currents should inevitably affect the state of local lakes in this situation, and respective events should be recorded in biological and lithological changes. Investigation of recent flora (Volkova *et al.*, 1995) revealed a higher species diversity in northern Mongolia than in adjacent regions. This seems to be related to the variability of ecological niches allowing plants to survive unfavorable periods in local refuges with the appropriate temperature and humidity. For instance, Malaeva (1989b) argued that the arboreal vegetation in the North Mongolian mountains survived even the most unfavorable periods of the Pleistocene, due to its vertical migration along the slopes with different exposure to light. She also concluded that forest areas in the North Mongolian mountains grew during the cold Pleistocene phases, because the lower forest boundary migrated downward. In view of the atmospheric precipitation deficit typical of the Asian interiors in general, we consider the lacustrine deposits as most suitable for pollen fossilization. Subsequently, studies on these deposits are most promising for the reconstruction of changes in the past environments, first of all, in the plant communities and climate. In the opinion of Krasheninnikov (1939), the present natural landscapes in northern Mongolia most resemble, among others in Eurasia, the "cold forest-steppe zones of the Pleistocene," and this opens a way of using the modern analogs by reconstructing the past.

PRESENT ENVIRONMENTS IN THE REGION

This work is based on data from lithological and biostratigraphic analyses of core samples of bottom sediments obtained from three lakes located in northern Mongolia between 50–52° N and 99–107° E inside the Hovsgol Highland (the Dood Nuur and Hovsgol lakes) and Orhon–Selenga depression (the Gun Nuur Lake; see Fig. 1).

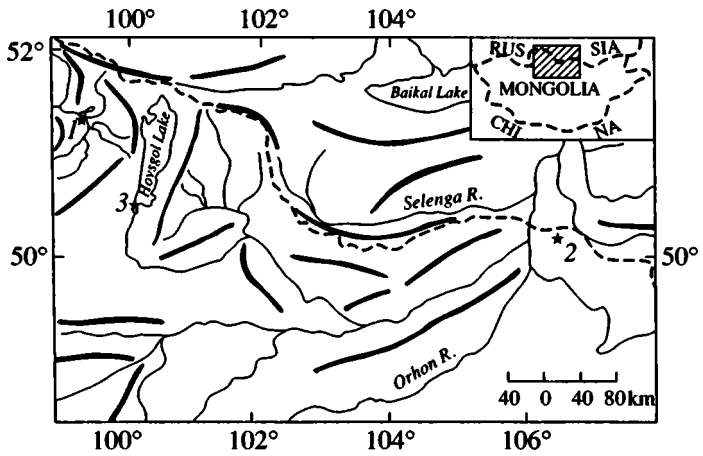

Fig. 1. Schematic map of the region illustrating position of lakes relative to the river system and mountain ridges: (*1*) Dood Nuur Lake (51°20′ N, 99°23′ E, altitude 1538 m); (*2*) Gun Nuur Lake (50°15′ N, 106°36′ E, altitude 600 m); (*3*) Hovsgol Lake (50°32′ N, 101°10′ E, altitude 1645 m); sampling localities are marked with asterisks.

The submeridional ridges of the Hovsgol Highland, attaining an absolute height of 3000–3200 m, separate two rift depressions of the Baikal type: the Hovsgol depression with the synonymous lake and the Darhat depression with the Dood Nuur Lake (Zolotarev *et al.*, 1982). The highland is incised by deep and narrow canyons. In the east of the region, a series of radiating ridges with a complex topography and altitudes below 1550 m surrounds the wide depression, filled with alluvial deposits of the Orhon and Selenga rivers (Fig. 1).

Marks of old Pleistocene glaciation are recorded in all mountain massifs of Mongolia (Sevast'yanov and Tserensodnom, 1994; Shuvalov and Nikolaeva, 1994). Glaciers are absent at present, however. Permafrost localities are widespread here as a consequence of severe low-snow winter seasons. These localities are marked by thermokarst relief forms, indicating hot summer seasons. In the northern part of the country, the permafrost horizons may occur at a depth of only 1–1.5 m (Sevast'yanov and Tserensodnom, 1994).

The differentiated topography, ranging from high mountains to flat elevated plains, and the isolation of the region from oceanic atmospheric masses are responsible for the distinctly continental climate of Mongolia, with a low degree of atmospheric precipitation and a high amplitude of daily and annual temperature fluctuations (Beresneva, 1995).

The Asiatic anticyclone, whose center is in northern Mongolia, is an important climate-forming factor during the winter periods. The windless weather with few clouds is favorable for radiation cooling, which causes a deep freezing of the ground. Stagnation of the cold and dry air in the intermontane depressions further stimulates a temperature decrease. In the north of the country, the average temperature during January is in the range of –23 to –27°C, whereas its peak value may be as low as –48°C (*Natsional'nyi...*, 1990).

The ground warming in the summertime is high. In the Hovsgol region, the average temperature of July is between 17 and 20°C, and can be as high as 32–35°C (Bufal *et al.*, 1989). These parameters for the Orhon–Selenga depression are 16 and 38.5°C, respectively (*Natsional'nyi...*, 1990).

Northern Mongolia is a rather humid country because of the mountains and general southeastern inclination of the territory toward the Pacific Ocean. About 60–70% of the annual precipitation falls here in the summer period, mainly in July and August (Beresneva, 1995), when the atmospheric pressure gradient between the heated continent and cooler ocean provokes a monsoon circulation, which brings the humid Pacific wind inside the continent and causes an intense rainfall in coastal regions of China and the Russian Far East. The systematic influence of the summer monsoon is typical of eastern Mongolia only (Beresneva, 1995). However, the monsoon may sometimes advance as far as the Transbaikal region and northern Mongolia, i.e., up to 100–107° E (Kornienko, 1975). At the altitude of about 1500 m, the eastern current of atmospheric masses is traceable up to 100° E (Beresneva, 1988). The amount of precipitation increases upward in the mountains, and, at the same time, its distribution significantly depends on the orientation of mountain ridges relative to the direction of atmospheric currents. In the Mongolian Altai, maximum humidity is typical of western mountain slopes, which are under influence, though weak, of the Atlantic Ocean (Beresneva, 1995). The maximum rainfall in northern Mongolia is up to 600 mm/year at the altitude of 2000 m (Bufal *et al.*, 1989). In coastal areas of the studied lakes, the annual rainfall is about 300–350 mm (*Natsional'nyi...*, 1990), whereas average evaporation may be as high as 400–550 mm/year (Kuznetsov, 1968).

The differentiated relief and a high range of topographic altitudes are responsible for the mosaic character of the vegetation cover's displaying a combination of latitudinal and vertical zoning. Principal zones correspond to the steppe, taiga, and bold peak vegetation types (*Karta...*, 1979; Batraeva *et al.*, 1989).

The studied lakes are natural components of the Enisei River system. The water balance in the lakes depends on the annual value of precipitation and summer temperatures (Sevast'yanov and Tserensodnom, 1994). The insufficient rainfall and hot summer periods decrease the lake levels and make the water mineralization easier, thus quickly affecting the composition of sediments, sedimentation rates, and evolution of organisms in the lakes. Small and shallow lakes of the steppe zone react to climatic changes more quickly than others. Data of instrumental monitoring show that fluctuations in humidity in the North Mongolian mountains are not as considerable as in the less elevated but more arid southern and eastern areas (Sevast'yanov and Tserensodnom, 1994).

CHARACTERISTICS OF THE LAKES

The Dood Nuur Lake (51°20′ N, 99°23′ E, 1538 m above sea level) is a relic of a vast lacustrine basin formerly spread over a considerable part of the Darhat depression during the pluvial Pleistocene epochs. As estimated on the basis of ancient terraces, the lake area attained 2600 km^2 at the altitude of 1700 m (Uflyand *et al.*, 1969). By the end of the late Pleistocene, its level significantly decreased, and the former lake was transformed into several basins connected by canals. The Dood Nuur Lake represents the most spacious of these basins and is 64 km^2 in area and –5...–7 m deep; its maximum depth is up to –15 m (Egorov *et al.*, 1994). The lake consists of semi-isolated Targan, Dund, and Harmay subbasins. The watershed area of rivers flowing in the Dood Nuur Lake is 1100 km^2. Most significant among them are the Shished Gol, Shargyn, and Arsayn Gol rivers. The first one serves also as the draining waterway.

The Dood Nuur Lake is an oligotrophic water basin. The content of mineral components in the water is 197 mg/l, and pH varies from 7.5 to 8.3.

Bottom sediments of the lake consist of coarse-grained material, gravel, sand, and mud. Sand to gravel facies are widespread above the isobath of 1.5 m, whereas the organogenic mud (gyttja) is typical of the central part of the lake (Sevast'yanov and Tserensodnom, 1994).

In the middle of July, shallow (4–5 m deep) zones near the coast are overgrown with macrophytes *Myriophyllum spicatum* L., *Ceratophyllum demersum* L., *Potamogeton pectinatus* L., *Persicoria amphibia* (L.) S.A., and *Chara* sp. (Dulmaa, 1965).

The phytoplankton of the lake includes 72 taxa, 37 of which are represented by diatomaceous algae (Dorofeyuk and Tsetsegmaa, 1994). *Asterionella formosa* Hass. is the dominant taxon of the summer season, whereas *Cyclotella* species are preponderant in the fall and winter periods. Bottom sediments here yielded 233 taxa of diatoms, mainly their benthic forms. The group of dominant species consists of *Paralia arenaria* (Moore) Moiss., *Fragilaria brevistriata* Grun., *F. construens* (Ehr.) Grun., *F. pinnata* Ehr., *Cocconeis placentula* Ehr., and *Amphora ovalis* Kutz.

The surrounding terrestrial vegetation represents a combination of forest and steppe coenoses (*Karta...*, 1979). Gently sloping eastern and northwestern coasts with abundant and periodically flooded hollows are covered with gramineous *Poa attenuata* Trin., *Festuca lenensis* Drob., *Koeleria cristata* (L.) Pres., and *Agropyron cristatum* (L.) associated with the grassland assemblage of *Carex enervis* C.A. Mey, *C. duriuscula* C.A. Mey, and *Kobresia myosuroides* (Vill.) Fiori et Paol. Toward the south, where the landscape is 5–12 m higher, the recreational larch forests (*Larix sibirica* Ledeb.) include isolated areas overgrown with Gramineae and steppe forbs: *Poa sibirica* Roshev., *Trisetum sibiricum* Rupr., *Ranunculus borealis* Trautv., and *Pulsatilla patens* (L.) Mill.

The Hovsgol Lake (50°32′ N, 101°10′ E, 1645 m above sea level) is in the central part of the graben extended in the meridional direction and formed in the second half of the Paleogene (Zolotarev *et al.*, 1989). Areas of the lake and its watershed are 2760 and 4920 km^2, respectively (Egorov *et al.*, 1994). In the north, the lake is bounded by the 50-meter-high escarpment of the Munku–Sardyk Ridge. In the west, it is surrounded by steep slopes of the Hordil–Sardag and Bayan–Uul ridges, whereas mountains of the eastern coast are lower and less steep. The southern coastline delineates the narrow Hatgal Bay. Three subaqueous and six overland terrace levels detectable in the lake depression were presumably formed in the middle–late Pleistocene and Holocene (Zolotarev *et al.*, 1989).

In addition to atmospheric precipitation, there are 46 creeks and rivers contributing to the lake's alimentation. The water discharge happens in the south through the Egiyn Gol River, the left-hand tributary of the Selenga River.

The Hovsgol Lake is deep: the average and maximum depth levels of the lake correspond to 138 and 262 m, respectively. Its littoral zone, shallower than 50 m, occupies about 15% of the basin area (Tomilov and Dashidorzh, 1965). The grain size of bottom sediments gradually decreases away from the coast simultaneously with a depth increase. Coarse-grained deposits and sands prevail in the coastal zone, whereas silty, pelitic, and clay muds are concentrated in the deeper central part of the lake. Carbonate mud is widespread in its southern and southeastern areas (Badrakh *et al.*, 1989). According to Tomilov and Dashidorzh (1965), the light-gray calcareous mud is characteristic of the depth interval of 3–25 m.

The macrophyte vegetation (chiefly *Potamogeton*) is confined only to shallow bays and river-mouth shoals, which are less than 6 m deep and covered with sandy mud (Tomilov and Dashidorzh, 1965). In littoral and sublittoral zones, the bottom is overgrown with large algae down to the depth of 50–60 m (Kozhova, 1983; Kozhova *et al.*, 1986).

The phytoplankton of the lake is represented by 55 taxa (Kozhova, 1983). Diatomaceous and chlorococcacean species are very diverse, and *Cyclotella ocellata* Pant. is annually dominant. Bottom deposits contain 202 diatom taxa, 19 of which are planktonic (Dorofeyuk, 1977). Preponderant forms of fossil diatoms are *Cyclotella ocellata, Opephora martyi* Herib., *Fragilaria brevistriata* Grun., *F. construens* (Ehr.) Grun. et var., *F. pinnata* Ehr., and *Amphora ovalis* Kutz. et var.

Vegetation around the Hovsgol Lake is distributed in a mosaic mode and clearly demonstrates vertical zoning (*Karta...*, 1979; Batraeva *et al.*, 1989). Lowered areas in the Egiyn–Gol valley are populated by steppe assemblages with preponderant *Stipa krylovii* Roshev. and *Agropyron cristatum* (L.) Beauv. Larch forests with an admixture of fruticose *Betula fruticosa* Pall., *Rhododendron parvifolium* Adams, and *Potontilla fruticosa* (L.) are typical of river terraces, where they alternate with steppe areas occupied by gramineous *Poa attenuata* Trin., *Koeleria cristata* (L.) Pers., and *Festuca lenensis* Drob. The mountain forest–steppe assemblages of this type are widespread within the altitudinal range of 1645–1900 m eastward and westward of the lake. The mountain forest belt (altitude of 1900–2200 m) consists of either larch forests with abundant fruticose and fruticulose species (*Rhododendron dauricum* L., *Vaccinium vitis-idaea* L., *V. uliginosum* L., *Ledum palustre* L.) or larch–pine forests (*Pinus sibirica* Du Tour., *Larix sibirica* Ledeb.) with abundant gramineae and forbs. A zone below the bold-peak belt (altitudes of 2200–2300 m) comprises light larch forests with *Betula rotundifolia* Spach., *Rhododendron parvifolium* Adams, and *Caragana jubata* (Pall.) Poir., and also light pine forests with a fruticose assemblage of *Rhododendron parvifolium* and *Rosa acicularis* Lindl. The bold-peak vegetation (2300–2800 m) consists of tundra taxa *Kobresia myosuroides* (Vill.) Fiori, *Dryas oxyodonta* Juz., and lichens (Batraeva *et al.*, 1989). Vegetation above the snow line is almost absent.

The Gun Nuur Lake (50°15′ N, 106°36′ E, 600 m above sea level) is situated in the east of the Orhon–Selenga depression. The lake terraces have relative altitudes of 6, 3–4, and 1.5 m and occur in the eastern, northern, and western coast sectors and mark the respective transgressive phases in the lake history. One more subaqueous terrace corresponding to a low lake level in the past is now located at a depth of 1.2 m. The lake area is 2.5 km^2; its average and maximum depth levels are 2 and 5 m, respectively. The Gun Nuur Lake is an eutrophic water basin with pH 8. Atmospheric precipitation and ground waters are the main alimentation sources of the lake. At present, water discharge happens periodically, and the lake is almost completely isolated from the Buryn Gol River, the right-hand tributary of the Orhon River. Marshy coasts of the lake are overgrown with *Phragmites* and *Scirpus*. Sand dunes, which are located south of the lake, are populated by *Pinus sylvestris* L., associated with accessory *Ulmus pumila* L. and *Salix*. Gramineae and forbs consisting of *Koeleria glauca* (*Spreng.*) D.C., *Agropyron cristatum, Festuca lenensis, Artemisia frigida* Willd., *Thermopsis lanceolata* R. Bz., and *Caragana microphylla* (Pall.) Lam. grow in the low relief forms of the depression (*Karta...*, 1979).

INVESTIGATION METHODS

Core samples for all the analyses performed were collected by B.P. Vipper and N.I. Dorofeyuk during the field work of the paleobotanical group of the Joint Biological Expedition of Mongolia and Academy of Sciences of the USSR. Sampling was performed from two boats by using either a piston coring (core catcher 22 cm long and 3.5 cm in diameter) to sample the upper semiliquid layer of sediments, or the "Russian corer" (core catcher 50 cm long and 6 cm in diameter) to sample more dense sediments.

Stratigraphic columns were described directly at the sampling sites. Samples for the diatom and pollen analyses were collected in the form of monolithic core portions 2–3 cm long, characterizing all changes in the lithological composition of deposits.

Chronological levels of the deposits are substantiated by the 11 radiocarbon dates obtained for three columns. These dates are insufficient for the reliable reconstruction of all secular changes in vegetation and lake levels. The regional palynostratigraphic scheme is also unsatisfactory so far. In this situation, we used the linear interpolation method to date sedimentary beds and boundaries of pollen zones between the measured radiocarbon dates.

Samples for diatom analysis were subjected to the standard treatment (*Diatomovye vodorosli...*, 1974). An abundance of diatom valves was calculated per one gram of a given rock sample (Davydova and Strel'nikova, 1979; Davydova, 1985). The alga types attributable to different salinity ranges were identified by using the standard halolimnetic scale (Kolbe, 1927; Proshkina-Lavrenko, 1953).

The pollen analysis of samples, which were prepared by the standard procedure with subsequent acetolysis (Grichuk and Zaklinskaya, 1948), was carried out by V.T. Sokolovskaya. In samples from the Gun Nuur and Hovsgol lakes, pine pollen was identified at the subgenus level. In the studied region, pollen of *Pinus* s/g *Diploxylon* can be produced only by the ordinary pine (*Pinus sylvestris* L.), whereas pollen of *Pinus* s/g *Haploxylon* represents the production of Siberian pine (*Pinus sibirica* Du Tour.). We did not discriminate

between the pine pollen in samples from the Dood Nuur Lake, and in the respective diagrams, this pollen may indicate both species.

When plotting the diatom and pollen diagrams, we calculated the percentage of each taxon relative to the sum of microfossil taken for 100%.

Our interpretation of fossil pollen spectra is based on comparison with data on the surface samples investigated in Mongolia (Malaeva, 1989a) and Kazakhstan (Chupina, 1971; Tarasov, 1992). In areas of combined steppe and forest coenoses, samples of the forest ground contain not less than 50% of arboreal and fruticose pollen. The larch pollen content in samples that were taken directly near the larch forest boundary is insignificant, because this pollen's poor aerodynamic properties make it unsuitable for proper fossilization (Malaeva, 1989a). Pollen of pine, spruce, and fir is better flying, and its presence in spectra may be a result of wind contamination if the presence is lower than 3%. This factor should be always considered in speculations on the past flora composition in the region of study.

The method of detecting secular variations in the lake level is based on both biostratigraphic and lithological data (Harrison and Digerfeldt, 1993). Earlier, it was successfully used in studies of lakes in the semiarid and humid zones of northern Eurasia, including some lacustrine basins of Mongolia (Tarasov *et al.*, 1994).

HISTORY OF VEGETATION EVOLUTION

Among the deposits we studied, the oldest are from the Dood Nuur Lake. The local pollen zone DN1 in the subbottom depth interval (SDI) of 265–270 cm shows here the maximum presence of arboreal pollen (up to 85% in sum), mainly represented by species of *Pinus* (40%), *Picea* (30%), *Larix* (up to 4%), and *Betula* (up to 10%) (Fig. 2). It also yielded the less abundant pollen of *Abies, Ulmus,* and *Alnus*. The associated herbaceous pollen consists chiefly of *Artemisia* and accessory Chenopodiaceae, Poaceae, and Cyperaceae specimens. This composition of pollen spectra suggests that forest areas of the respective time were more widespread in the Darhat depression than at present. Spruce and larch were principal forest plants at that time and coexisted with mixed birch–pine assemblages in the coastal zone of the lake. Our extrapolation of sedimentation rates by using the radiocarbon date of 11 470 ± 100 years (Vib-113) located higher in the section allows us to conventionally refer this phase to the terminal late Pleistocene (12 200–12 500 years ago).

The next zone, DN2 (SDI 205–265 cm), is marked by a considerably decreased content (35–45%) of arboreal pollen; specimens of *Picea* (5–10%) and *Pinus* (10%) become less abundant here, whereas pollen of *Ulmus* and *Abies* is absent. The herbaceous pollen shows a greater abundance of *Artemisia* (25–45%), Compositae (up to 10%), Chenopodiaceae (5–10%), and Poaceae (5–10%) species. Rosaceae and *Ephedra* are the new taxa to appear at this level. Changes in the pollen spectra suggest a lower abundance of dark conifers in the forest vegetation and an expansion of steppe areas in response to a humidity decrease and higher annual amplitudes of the air temperature. At this time (10 600–12 200 years ago), the hygrophyte arboreal taxa presumably migrated at higher hypsometric levels. The light larch forests and birch groves less sensitive to the deficient humidity could coexist well with petrophyte steppe vegetation near the lake.

Zone DN3 (SDI 0–205 cm) is again quite rich in arboreal pollen (60–80%). However, the abundance of *Picea* specimens further decreases here, up to their complete disappearance in the uppermost sample. *Pinus* and *Betula* species remain dominant and are associated with *Larix* and *Alnus* pollen, whose content is always between 1 and 5%, and also with sporadic pollen grains of *Abies* and *Ulmus*. Herbaceous components of the spectra are represented by *Artemisia* (10–20%) and the less abundant Chenopodiaceae, Poaceae, Ranunculaceae, Brassicaceae, and Compositae, as it is typical of the grassland steppe coenoses. Against the background of constant composition of herbaceous plants, there is a notable decrease in the abundance of *Pinus* pollen down to 25% and a simultaneous growth of the *Betula* content up to 45% at the depth of 0.7 m. The same level marks the extinction of *Ulmus* pollen and a minor peak in the abundance of *Abies* specimens. Interpreting these data, we assume that proportions between the forest and steppe areas near the Dood Nuur Lake were rather stable during the entire Holocene interval. The total area of forests with dominant pine, birch, and larch trees here was nearly the same or even larger than at present. The relatively low content (60%) of arboreal pollen in the uppermost bottom sediments is probably a result of the anthropogenic influence on the landscape, however, this should be substantiated by additional data.

The recovered Gun Nuur section spans only the Holocene interval of deposits (Fig. 3).

The basal zone GN1 of this section (SDI 490–493 cm) reveals approximately equal proportions of arboreal and herbaceous pollen (Fig. 3). Its dominant representatives are *Betula* (30%), *Pinus* s/g *Diploxylon* (about 20%), Chenopodiaceae (20%), and *Artemisia* (15%). Pollen of this composition may correspond to the *Chenopodium–Artemisia* pioneer vegetation formerly widespread in the lake depression. Pine trees (most likely *Pinus sylvestris* as at present) grew in limited areas of sand dunes, whereas birch groves preferred the more humid places near the lake. Extrapolation data, which are based on the radiocarbon date of 9550 ± 150 years (TA-1349A) for the higher section level, allow this phase to be dated by the age interval of 9600–10 000 years ago.

Zone GN2 (SDI 450–490 cm) reveals a distinct decrease in the presence of arboreal pollen (25–35%) mainly consisting of a *Betula* fraction mixed with an

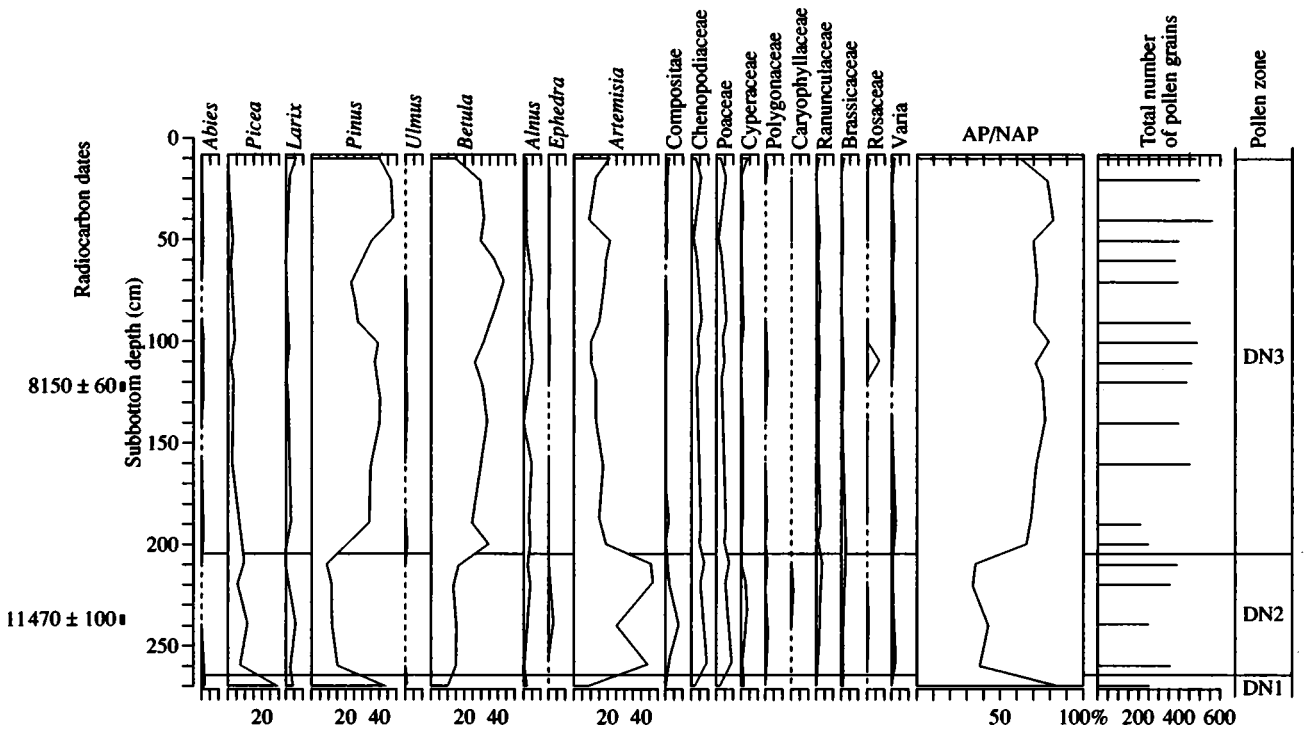

Fig. 2. Pollen diagram for sedimentary column of the Dood Nuur Lake (in all diagrams, AP/NAP indicates proportions between arboreal and herbaceous pollen in the spectra).

insignificant amount of pollen grains of conifers (*Abies, Picea, Pinus* s/g *Haploxylon*) and *Alnus fruticosa* Rupr. Pollen of *Artemisia* (20–35%), Chenopodiaceae (20–40%), Poaceae (5–10%), Cyperaceae (5–7%). The less abundant *Ephedra* is dominant in herbaceous spectra. The arboreal vegetation in the lake depression consisted mostly of birch groves with fruticose alder undergrowth. Judging from the low abundance of pine pollen, trees producing it grew away from the lake. Preponderant near the coast were the herbaceous communities of steppe, or dry steppe landscapes. On the basis of two radiocarbon dates, this phase can be confidently referred to the early Holocene (8700–9600 years ago).

Zone GN3 (SDI 300–450 cm) corresponds to the interval where the content of arboreal pollen gradually grows up to 60%. Upward the section, the initially dominant *Betula* species give their place to pollen of *Pinus* s/g *Diploxylon*, and this trend is accompanied by the appearance of specimens of *Ulmus, Larix, Abies*, and *Pinus* s/g *Haploxylon*. Pollen of the last two taxa most likely represents the result of wind contamination. As Poaceae (10–15%) and mesophilic forbs (5–10%) become more abundant in this zone, the content of Chenopodiaceae decreases (up to 10–15%), but *Artemisia* (10–20%) is still the dominant component. These pollen spectra may indicate that pine–birch forest "islands" in the lake valley included an admixture of larch and elm trees. At present, the latter grow in close proximity to the lake, but surprisingly, we found that elm pollen is absent in the surface layer of bottom sediments, and this may imply the considerable abundance of these plants in the past, about 8000–8700 years ago.

Zone GN4 (SDI 0–300 cm) yielded pollen assemblages compatible with current vegetation. In this case, the content of arboreal pollen is as high as 65–85%, and the dominant taxon is *Pinus* s/g *Diploxylon* (50–75%). Nearly all the samples studied also contain the accessory pollen of *Picea, Larix,* and *Alnus fruticosa.* Specimens of *Ulmus* disappear from spectra beginning with the depth of 250 cm. *Artemisia,* Poaceae, and Chenopodiaceae are dominant among herbaceous taxa, as before. A certain similarity between the subsurface and deeper pollen spectra from zone GN3 implies that the middle–late Holocene vegetation around the Gun Nuur Lake was of the current type.

Pollen analysis of samples from the Hovsgol Lake characterizes only the upper half of the Holocene (Fig. 4). The diagram reveals two pollen zones in this locality.

Zone H1 (SDI 140–220 cm) is very rich in arboreal pollen (90–95%), and specimens of *Pinus* s/g *Diploxylon* and *Pinus* s/g *Haploxylon* are dominant in its composition. Other taxa, such as *Larix* (1–5%), *Betula* (1–5%), *Alnus, Picea,* and *Abies,* are accessory. *Artemisia* (up to 7%) preponderates in herbaceous group also comprising petrophytes (Chenopodiaceae, *Ephedra*) and forest–grassland taxa (Poaceae, Ranunculaceae). Compositae and Caryophyllaceae, characteristic of steppe coenoses, are absent. The data show that pine

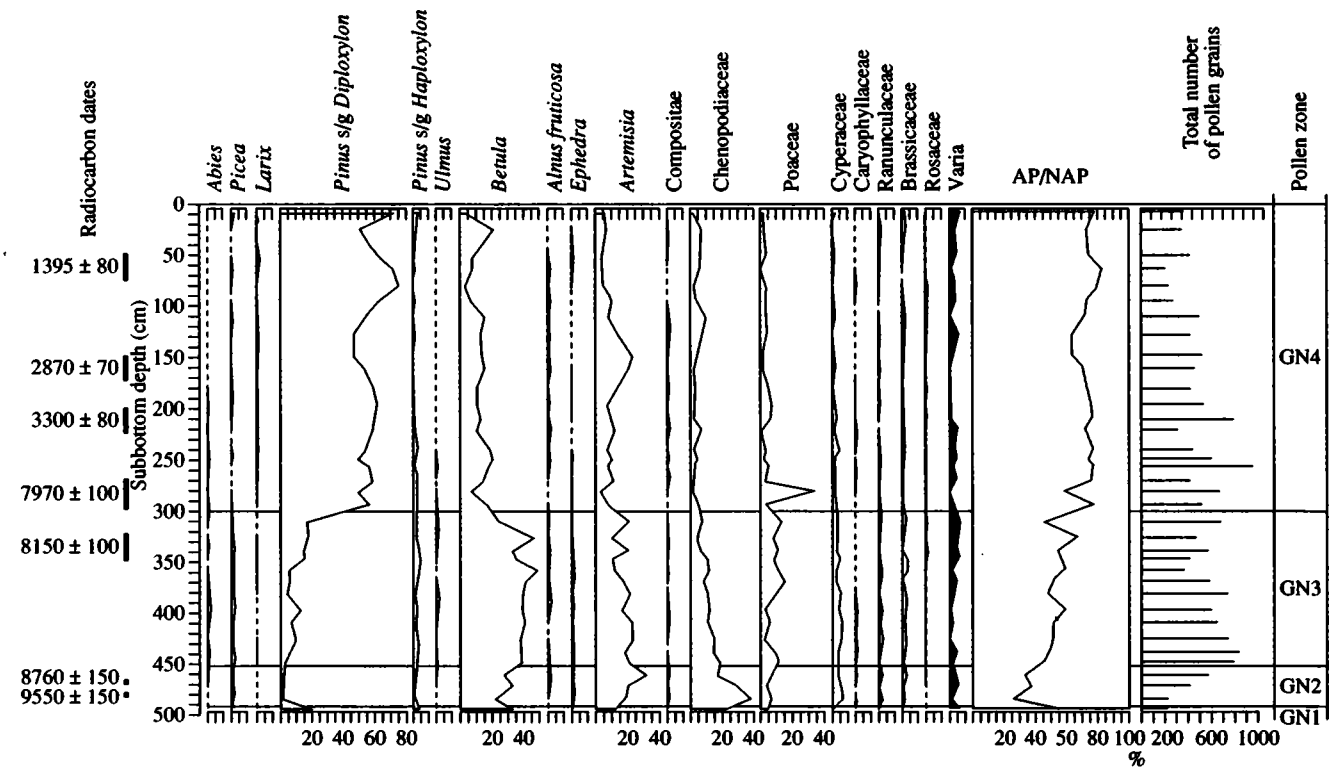

Fig. 3. Pollen diagram for sedimentary column of the Gun Nuur Lake.

forests with accessory larch and spruce trees were widespread in the lake depression 4300–6350 years ago. This age interval is well substantiated by two radiocarbon dates of 5800 ± 100 (TA-671) and 3910 ± 60 (TA-670) years (Fig. 4).

Zone H2 (SDI 0–140 cm) shows a slightly lower content of arboreal pollen (75–90%). The still dominant *Pinus* s/g *Diploxylon* and *Pinus* s/g *Haploxylon* associate here with more abundant *Larix* (up to 10–17%) and *Betula* (up to 5–7%). *Picea* is another permanent component, but *Abies* disappears at the depth of 85 cm below the bottom surface (about 3100 years ago). The pollen fraction of herbaceous plants is rather rich in *Artemisia* (10–17%) and also includes Compositae, Caryophyllaceae, and Brassicaceae, suggesting a greater significance of steppe coenoses. This vegetation was quite close to the recent flora.

SECULAR VARIATIONS IN THE LAKE LEVELS

The Dood Nuur Lake. The coring site at a depth of 3.3 m in the southeastern part of the Harmay Bay was remote for 200 m from the coast and penetrated through the 270-cm-thick sequence of bottom sediments. Changes in the lake level, beginning from the terminal late Pleistocene, are depicted on the basis of lithologic and biostratigraphic data.

The basal sand of the sequence (SDI 265–270 cm) is free of fossil diatoms (Fig. 5). This layer was probably deposited in the mouth area of the past water channel about 12 300–12 500 years ago, when the lake level was considerably lower than at present.

Sandy clay with detrital plants remains, and molluscan shells from SDI 227–265 cm presumably marks the incipient transgression phase. This unit yielded 54 diatom taxa, among which there were dominant *Paralia arenaria* (Moore) Moiss. (26.1%) and *P. scabrosa* (Oestr.) Moiss. (14.8%), the benthic forms populating the well-aerated sandy shoals, and also the epiphyte *Fragilaria pinnata*. In the interval of 240–265 cm, the content of planktonic diatoms is below 6%, but higher, in the horizon of 227–240 cm, it is as high as 17%. In the last case, the dominant form is *Cyclotella ocellata*, implying the gradual rising of the lake level. The radiocarbon age of organic remains from the depth of 235–240 cm is 11 470 ± 100 years (Vib-113), and the transgressive phase in question is dated at 11 200–12 300 years ago.

The sandy clay bed grades upward into the sandy–clayey gyttja (SDI 190–227 cm), containing plant remains and molluscan shells and suggesting the relative shoaling of the lake. Such a conclusion is also inferable from the lower concentration and diversity of diatoms at this level (only 24 taxa), where they demonstrate the low abundance of planktonic forms and absolute preponderance of benthic species *Paralia arenaria* and epiphytes. According to our calculation, this regressive phase took place 10 100–11 200 years ago.

Overlying sediments (SDI 53–190 cm) are represented by clayey gyttja (Fig. 5). We believe that the sand fraction disappearance at this level indicates a deep-water sedimentation environment. Compositional variations in diatom assemblages mark the lower-rank fluctuations of the lake level. At first, in the depth interval of 120–190 cm, planktonic diatoms become rather abundant (21% at the depth of 150 cm) due to the lake-level rise about 9000 years ago, and then their content decreases down to the initial level. The constant presence of plant remains and the general preponderance of benthic diatoms and epiphytes in association with planktonic *Aulacosira ambigua* do not allow us to consider the inferable fluctuations of the lake level as significant. The radiocarbon age (8150 ± 60 years; Vib-112) of organic remains from the depth interval of 120–125 cm allows the respective phase of sedimentation to be dated as far back as 8000 years ago.

The diatom assemblage from the clayey gyttja bed at the depth of 90–120 cm consists of 113 taxa and thus marks the period of their thriving. The content of planktonic forms is low at this level (2–6.7%). It appears that the lake level was not very high, and the basin was mesotrophic and overgrown with macrophytes 6000–8000 years ago.

The next level, at the depth of 77–90 cm, recorded considerable changes in the diatom population: their presence and species diversity (18 taxa) abruptly decreased here: planktonic forms ceased to exist, and benthic *Paralia arenaria* (86.5%) turned into the principal dominant component. These changes are indicators of the further shoaling of the basin 5200–6000 years ago.

The diatom assemblage of clayey gyttja at the depth of 53–77 cm corresponds to the water-rich phase in the lake's history, as it follows from the abundance of planktonic species grown up to 21%. This transgressive phase arose 3600–5200 years ago.

The upper sandy gyttja bed (SDI 0–53 cm) with undecomposed plant remains and molluscan shells was deposited during the last 3600 years in a shallow basin, the littoral zones of which were overgrown with macrophytes. This conclusion is inferred from the low content of planktonic diatoms (5%) and the high abundance of epiphytes (up to 85%).

The Gun Nuur Lake. In this case, our inferences on the lake-level changes are also based on the lithological features of bottom sediments and on the data of diatom analysis (Fig. 6).

The sedimentary sequence was penetrated by drilling in the southeastern part of the lake 100 m away from the coast, where the depth is 4.5 m. At the base of the sequence (490–493 cm), there is a bed of coarse-grained sand with an insignificant admixture of silty material, suggesting that sedimentation proceeded under a low lake level. The absence of diatoms in this bed also indicates shallow-water environments. The next bed of calcareous clay with an admixture of a coarse-grained fraction seems to have been deposited when the lake level was rising. At the same time, several features, such as the poor sorting of sedimentary material, the high abundance of Cyperaceae pollen, and the scarcity of diatom valves in this bed, show that the rise was not significant. Radiocarbon dates of 9550 ± 150 and 8760 ± 150 years were obtained for two sam-

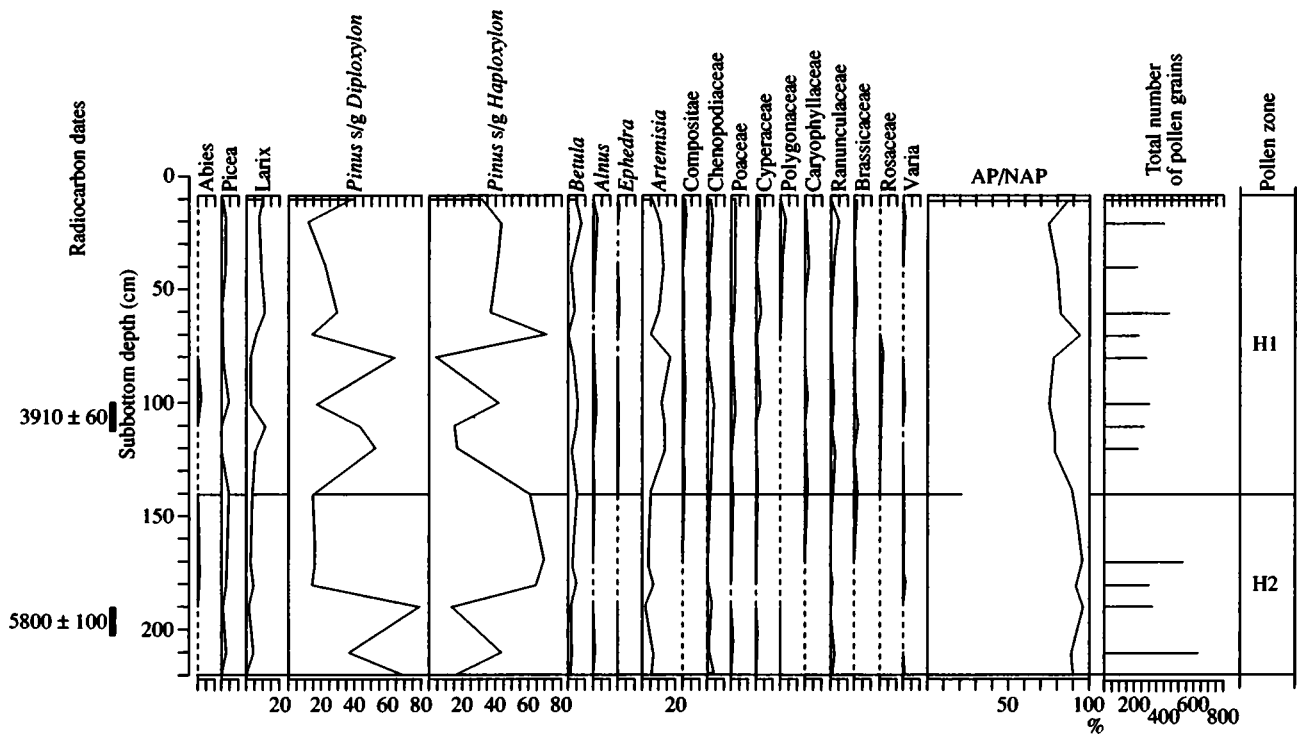

Fig. 4. Pollen diagram for sedimentary column of the Hovsgol Lake.

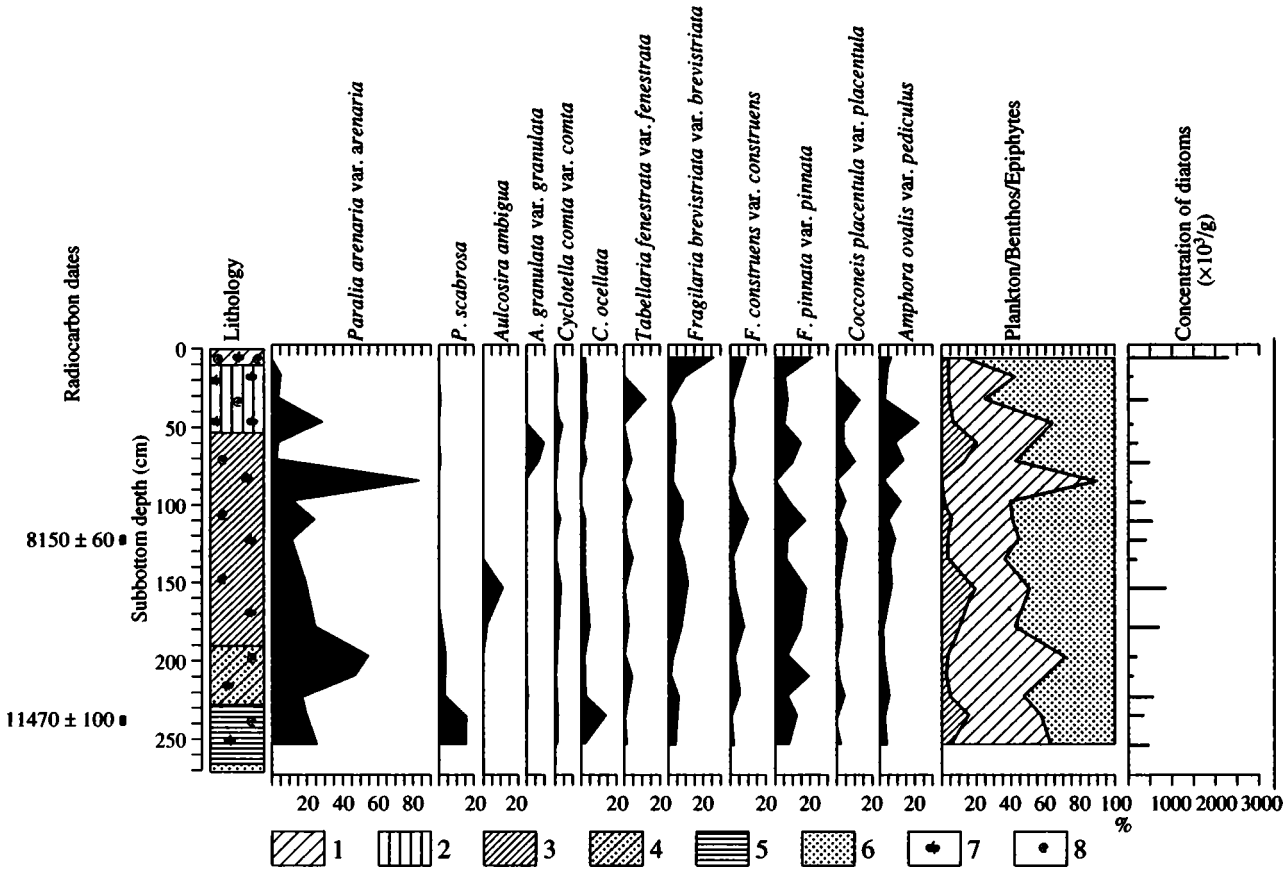

Fig. 5. Lithology and simplified pollen diagram for the sedimentary column of the Dood Nuur Lake: (1) sandy gyttja; (2) gyttja; (3) clayey gyttja; (4) sandy-clayey gyttja; (5) sandy clay; (6) sand; (7) plant remains; (8) molluscan shells.

ples of organic matter (TA-1349A and TA-1348A, respectively) from the lower part of the sequence, in order to exclude the overestimated age values possible by measuring them in the carbonate rocks. These dates allowed us to calculate that sediments at the depth of 480–493 cm were accumulated 9300–10 000 years ago.

The overlying bed (SDI 353–480 cm) is composed of marl. When the climate humidity is deficient, a intense calcium carbonate precipitation begins sumultaneously with a water-temperature rise, and, subsequently, this interval should correspond to a low lake level. Shallow-water environments in the lake at that time are also indicated by the relatively high abundance of Cyperaceae pollen and the extremely low content of diatoms represented by benthic forms. Radiocarbon age measurements date this phase back to 8300–9300 years ago.

The clayey gyttja bed, with an accessory fraction of fine-grained sand (SDI 246–353 cm), was deposited at a greater depth as compared to the current bottom level. In addition to the bed lithology, this is evident from the grown abundance and taxonomic diversity of diatoms, and also from the minimum content of Cyperaceae pollen. In the lower part of the bed (SDI 290–353 cm), diatoms are still not abundant and are represented by benthic *Amphora ovalis* var. *ovalis* Kutz. and *Anomoeoneis sphaerophora* var. *polygramma* (Ehr.) O. Mull. The last taxon is mesohalolimnetic and points to the relatively high mineralization of the lake water, which was probably inherited from the previous stage of the lake's evolution. Both taxa coexist with planktonic *Stephanodiscus rotula* var. *minutulus* (Kutz.) Hendey. At the depth of 270–280 cm, mesohalolimnetic forms disappear, and diatoms here are very rich in planktonic forms (90–95%), probably in response to the lake-level rise to the maximum. Radiocarbon dates of 8150 ± 100 (TA-1350A) and 7970 ± 100 years (TA-1417A) allow us to refer this phase to the middle Holocene (5400–8300 years ago).

Molluscan shells from the clayey and sandy gyttja bed (SDI 218–246 cm), where Cyperaceae pollen simultaneously grows in abundance, suggest somewhat shallower conditions of sedimentation in the lake. The taxonomic composition of diatoms is also in agreement with this inference, because their planktonic forms are somewhat less abundant here (35% at the boundary with the underlying bed), whereas benthic species are more numerous. At the same time, mesohalolimnetic forms are missing in this bed, and the abundance of planktonic diatoms quickly moves to the normal level. Therefore, the inferable decrease of the lake level

apparently was episodic, and the basin depth generally remained considerable 3300 –5400 years ago.

In the sandy gyttja bed at a depth of 108–218 cm, the diatom concentration drops. The abundance of planktonic forms here is lesser than 5–10%, and mesohalolimnetic *Anomoeoneis sphaerophora* var. *polygramma* appears once more among the dominant species. This diatom assemblage reflects a distinct phase of the lake shoaling and higher water mineralization. Radiocarbon dates measured in this bed are equal to 3300 ± 80 (TA-1482A) and 2870 ± 70 years (TA-1481A), and the corresponding regressive phase is dated back to about 2100–3300 years ago.

Undecomposed plant remains and small-size gravel appearing in the gyttja bed at the depth of 60–108 cm might indicate the further lowering of the lake level. However, characteristics of diatom assemblages from this interval contradict this assumption. The concentration of diatoms considerably rises here, and their planktonic forms become as abundant as 10–15%, whereas mesohalolimnetic species become rare. These changes are indicators of the water inflow. It seems that gravel appeared in the lake when it flooded land areas in the coastal zone 1300–2100 years ago. The upper boundary of the bed is reliably dated on the basis of radiocarbon age value at 1395 ± 80 years (TA-1441).

The upper gyttja bed, with a fine sand admixture (SDI 0–60 cm), seems to be accumulated at the shallower depth close to its present level. Diatoms are much less abundant here; the content of their planktonic forms is not higher than a few percent, but the concentration of mesohalolimnetic taxa rises up to 45%.

The Hovsgol Lake. Fluctuations of the lake level are described on the basis of lithology and diatom distribution in bottom sediments sampled in the northern part of the Hatgal Bay, 200 m away from its western coast (Fig. 7). The bottom depth here is 2.8 m.

The basal sandy-gravel deposits of the sequence (SDI 210–220 cm) are barren of fossils. The lithological features of these deposits indicate their accumulation in a subaerial setting when the lake level was very low.

The subsequent humidification of environments and the lake-level rise resulted about 6000–6100 years ago in the deposition of silty sand with accessory gravel-size clastics. From the depth interval of 205–210 cm, we recovered an impoverished diatom flora consisting of 37 taxa. Dominant among them are epiphyte *Fragilaria pinnata* (36.7%) and *Opephora martyi* (35.2%),

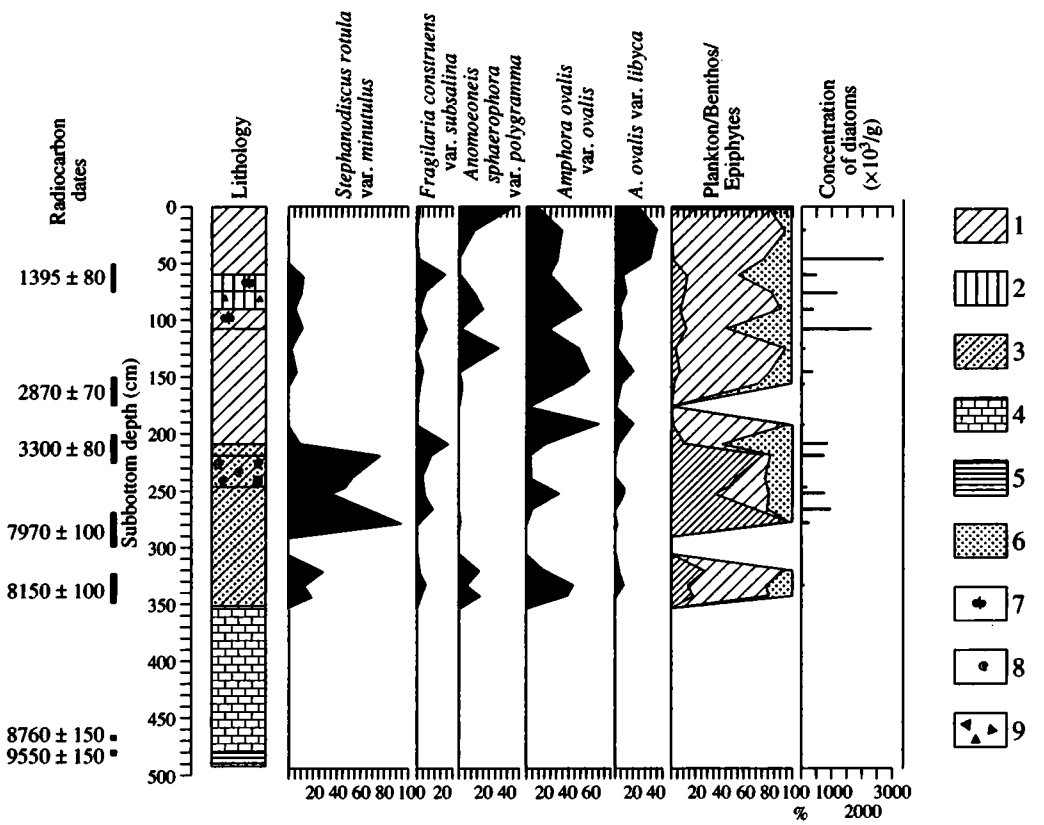

Fig. 6. Lithology and simplified pollen diagram for sedimentary column of the Gun Nuur Lake: (1) sandy gyttja; (2) gyttja; (3) sandy-clayey gyttja; (4) marl; (5) calcareous sandy clay; (6) coarse-grained sand; (7) plant remains; (8) molluscan shells; (9) gravel.

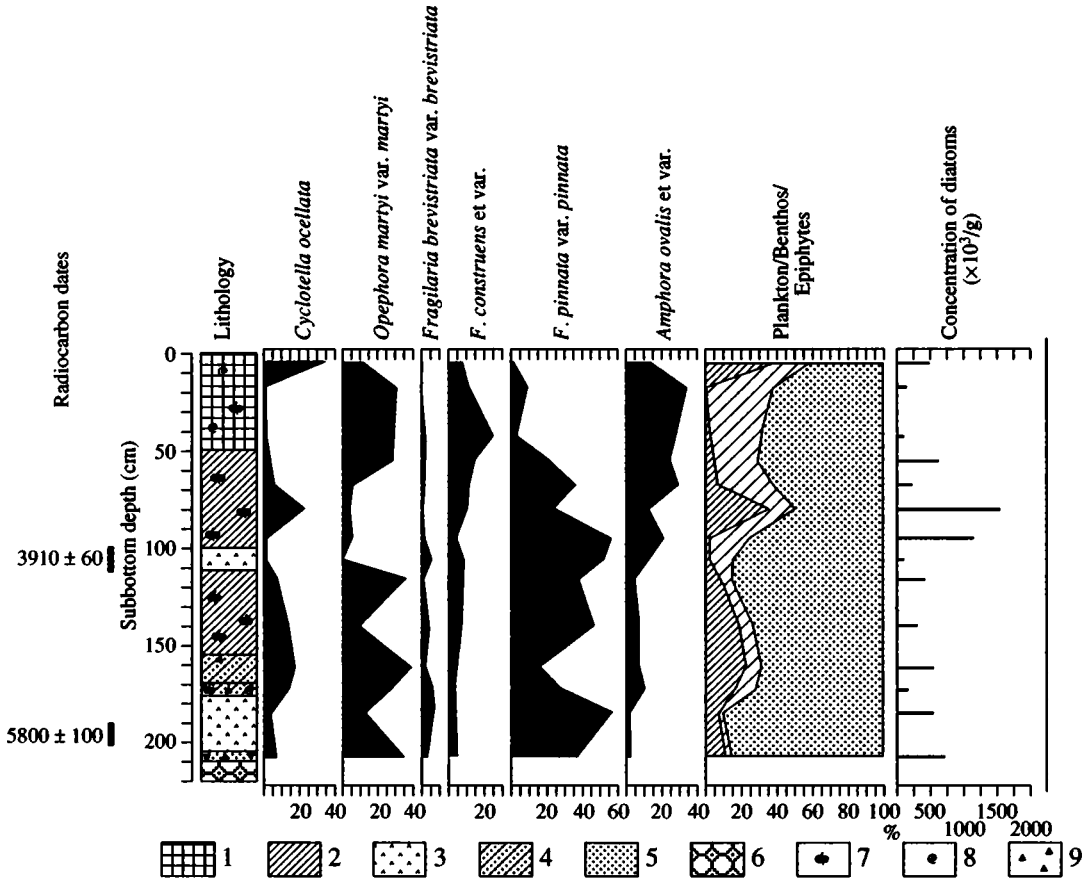

Fig. 7. Lithology and simplified pollen diagram for sedimentary column of the Hovsgol Lake: (1) calcareous gyttja; (2) clayey gyttja; (3) peat; (4) sandy-clayey gyttja; (5) sand; (6) unsorted sand with gravel; (7) plant remains; (8) molluscan shells; (9) gravel.

whereas planktonic *Cyclotella ocellata* (6%) is notably less abundant.

The next short-term shoaling is marked by the accumulation of a peat interlayer (SDI 176–205 cm) containing wood fragments and the remains of herbaceous plants. Diatoms from this interval are mainly represented by epiphytes (91%). The radiocarbon age of the peat sample (SDI 190–202 cm) is 5800 ± 100 years (TA-671).

The new lake-level rise is indicated by deposition of sandy-clayey gyttja above the peat bed (SDI 155–176 cm). In the interval of 170–176 cm, gyttja includes plant remains, molluscan shells, and small-size gravel. These lithological features suggest a quick transgression into the lake depression. Planktonic diatoms grew here in abundance up to 22%, and one of them, namely *Cyclotella ocellata,* moved in the dominant group. We calculated that this phase took place 4900–5400 years ago.

The clayey gyttja bed (SDI 112–155 cm) is rich in plant remains and reveals the gradual decline in abundance (down to 8%) of planktonic diatoms. Epiphyte forms preponderant in this interval are characteristic of a lake shoal overgrown with macrophytes. The calculated age interval of this regressive phase is 4000–4900 years.

The culmination of the lake-level drop was marked by deposition of the next peat bed (SDI 100–112 cm). The abundance of planktonic diatoms decreases here down to 2%, and dominant taxa of this flora are exemplified by epiphytes (*Fragilaria pinnata, F. construens* et var., and *F. brevistriata*) and benthic forms (*Amphora ovalis* var. *pediculus*). The radiocarbon age measurement for the peat sample yielded the value of 3910 ± 60 years (TA-670).

Above the peat, there is a clayey gyttja bed (SDI 70–100 cm) that recorded the new rise of the lake level 2600–3650 years ago. Diatoms turn here to be quite abundant and diverse. The content of their planktonic forms increases up to 35%, and two planktonic forms, *Cyclotella ocellata* and *C. kuetzingiana* var. *radiosa* Fricke, are components of the dominant group. However, the overlying gyttja bed (SDI 50–70 cm) marks another phase of sedimentation that took place 1800–2600 years ago. Concentration of diatoms decreases at this level by seven times, and the content of their planktonic forms is lower by a factor of five. These indicators imply the deterioration of dwelling conditions for diatoms because of the lake shoaling.

In the next bed of calcareous gyttja containing plant remains and molluscan shells (SDI 10–50 cm), we reg-

istered another drop in the abundance of diatoms, whose planktonic forms constitute 0.5% of the assemblage. Species dominant here are *Opephora martyi* var. *martyi, Amphora ovalis* var. *pediculus, Fragilaria pinnata,* and *F. construens* var. *venter* (Ehr.) Grun. These diatoms are characteristic of shallow-water settings with oligotrophic conditions, which have existed at the site about 350–1800 years ago.

The surface layer of carbonate gyttja shows an abrupt increase in the diatom abundance (37%). Their dominant group includes *Cyclotella ocellata, Amphora ovalis* var. *pediculus,* and *Opephora martyi.* This composition of diatom flora may imply a somewhat higher stand of the present lake level.

DISCUSSION

Available data on the history of the late Pleistocene–Holocene vegetation in northern Mongolia are inadequate and contradictory. Malaeva (1989b), who analyzed data pollen spectra in the region, noted several methodical errors: the absence of data on the surface samples, the insufficient frequency of sampling, and the statistically doubtful interpretation of pollen spectra. Many published speculations are based on results of pollen investigation in deposits of fluvial and lacustrine terraces, where one may anticipate stratigraphic breaks and the redeposition of fossil pollen. Radiocarbon dating is missing in many cases, and this complicates, or even makes impossible, the understanding and correlation of chronological phases in flora evolution detected by different investigators. Our analytical data on bottom deposits in three lakes offer a safe possibility to outline the vegetation evolution in northern Mongolia for the last 12 500 years, and to correlate the results with available conclusions on adjacent regions.

The composite pollen diagram for lacustrine deposits in central Mongolia (Vipper *et al.,* 1989) demonstrates that the forest phase, which is approximately 12 000–14 000 years old, has been replaced by the phase of steppe coenoses responsible for the peak abundance of herbaceous pollen 8000–11 000 years ago. In the last period, arboreal taxa were chiefly represented by *Betula* species, whereas pollen of *Picea* and *Pinus,* the main forest-forming plants of the late Pleistocene, was almost absent. The percentage of arboreal pollen in the spectra grew again in the Holocene time mainly due to the progressive development of pine forests. Its abundance was at the maximum about 6000 years ago (Vipper *et al.,* 1989), but somewhat lower in general than during the late Pleistocene. Spruce trees were not abundant in the early Holocene forests, as it follows from the low percentage of their pollen in the spectra. The late Holocene is again marked by progradation of herbaceous coenoses in central Mongolia. According to our data, the early Holocene phase (8700–9600 years ago) of degrading forests and prograding steppe landscapes took place around the Gun Nuur Lake, but was missing in the Darhat depression (the Dood Nuur Lake) located higher in the mountains. This inference is partially compatible with results obtained by Malaeva (1989b), who concluded that the "cold and arid boreal phase" is not detectable on the basis of pollen distribution in alluvial deposits of northern Mongolia. The asynchronism of natural events in different altitudinal zones of Mongolia is also reveled by instrumental methods of observations (Sevast'yanov and Tserensodnom, 1994). For instance, as the mountain lakes experienced transgression in the 1980s, the lake levels in plains were decreasing.

We feel it well grounded to state, especially on the basis of the reconstructed lake levels, that a similar asynchronism of events in the mountain and plain areas is one of characteristic features of the past. After the phase of the relatively high water level in the terminal late Pleistocene, the lakes of central Mongolia experienced a considerable shoaling about 9500 years ago (Vipper *et al.,* 1989; Tarasov *et al.,* 1994, 1995). During the Holocene, the lake levels were at the maximum about 7000–7500 years ago, and later, they were approaching or somewhat higher than the present state until the period of 3000–4000 years ago, when there was another phase of shoaling. A new rise of the lake levels happened 2000–2500 years ago (Tarasov *et al.,* 1995), and afterward, hydrological conditions turned to be close to the current situation. Secular level changes in the Gun Nuur Lake resemble those in the central Mongolian lakes. However, the mountain lakes of northern Mongolia developed in a different manner. After the late Pleistocene transgressive phase (11 200–12 300 years ago), the Dood Nuur Lake suffered shoaling 10 100–11 200 years ago. In contrast, both the Dood Nuur and Hovsgol lakes were at a low level during the middle Holocene. Their levels turned to rise again in the second half of the Holocene. The climatic dependence of these events is undoubtable. One of the works (Vipper *et al.,* 1989) explains these events as dependent on the fluctuations of the cyclone activity and summer temperature, and the middle Holocene phase of high levels in the central Mongolian lakes could be related to the activated monsoon circulation in response to the greater thermal and barometric gradients of that period between the cooler ocean and warmer land (Tarasov *et al.,* 1995). However, we should drop further speculations about the mechanism and driving forces responsible for climatic and vegetation changes in Mongolia during the late Pleistocene and Holocene, as they need to be substantiated by special study.

Paleogeographic information available for northern China allows us to conclude that the interval between 8000 and 3000 years ago was the most humid and warm during the last 18 000 years, and the monsoon circulation of that time was quite intense in the region. In the Beijing surroundings, the progradation of coniferous forests into the former tundra–steppe landscape of the late Pleistocene time was recorded 11 400–13 000 years ago (Liu, 1988). After the short-term dry period with progressing steppe coenoses (11 000–11 400 years ago),

forest formations turned again to be dominant here over the entire Holocene, and the most humid conditions appeared in the first half of the middle Holocene (Liu, 1988). A similar trend in flora evolution (Tarasov and Kremenetskii, 1995) and lake-level changes (Tarasov *et al.*, 1995) was also detected in the steppe zone of Kazakhstan. For instance, the pine-forest "islands" were most widespread in northern Kazakhstan during the second half of the Holocene, whereas birch was a principal forest-forming plant 6000–10 000 years ago. Lakes of this region attained their maximum level 6500–8000 years ago.

CONCLUSION

Palynological data for the Dood Nuur Lake show that forest vegetation was widespread in the Darhat depression 12 300–12 500 years ago, when spruce trees were more abundant. In central Mongolia, the same situation was reconstructed for the period of 12 000–14 000 years ago. Thus, the lower boundary of taiga forests in northern and central Mongolia was apparently lower at that time than at present.

The progradation of steppe vegetation was recorded the terminal Pleistocene (10 600–12 300 years ago) in the Darhat depression and at the beginning of the Holocene (8700–9600 years ago) in the Gun Nuur valley. After these relatively short events, forest areas increased again, and pine trees turned to be widespread in the studied regions. In the middle and late Holocene, the composition of forest vegetation and proportions between forest and steppe areas were much less variable.

In contrast to the vegetation cover, levels of the northern Mongolian lakes significantly fluctuated during the last 12 500 years, and respective data provide an important information about changes in the regional water budget. Since the arboreal vegetation is found to be rather insensitive with respect to humidity fluctuations (especially in mountain regions), we believe that secular temperature variations during the period in question were primarily responsible for the forest flora evolution.

There is a distinct asynchronism of humid phases in the mountain and plain areas of northern Mongolia. This is well illustrated by data on the lake-level changes. The high middle Holocene level recorded in the Gun Nuur Lake (altitude 600 m) is missing in the Dood Nuur Lake (altitude 1538 m). The middle Holocene phase of the lowermost lake levels took place in both the Hovsgol and Darhat depressions, when the Gun Nuur Lake experienced maximum transgression.

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Reviewer M.N. Alekseev

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BRIEF COMMUNICATIONS

First Discovery of Late Paleozoic Radiolarians in the Eastern Yakutiya

V. S. Rudenko*†, A. V. Prokop'ev, V. S. Oksman**, I. V. Kemkin*, and N. Yu. Bragin*****

** Far East Geological Institute, Far East Division, Russian Academy of Sciences, pr. Stoletiya Vladivostoka 159, Vladivostok, 690022 Russia*
*** Yakut Institute of Geological Sciences, Siberian Division, Russian Academy of Sciences, pr. Lenina 39, Yakutsk, 677891 Sakha Republic (Yakutiya)*
**** Geological Institute, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia*

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Abstract—Some assemblages of Late Paleozoic radiolarians are first established in Yakutiya. The radiolarians occur in hemipelagic deposits reduced in thickness. They are extracted from siliceous tuffites by using the hydrofluoric acid and studied under a scanning electron microscope. In total, five radiolarian assemblages are detected: two Early Carboniferous assemblages with *Albaillella* cf. *paradoxa* and *Albaillella indensis*, two Early Permian assemblages with *Latentifistula astricta* and *Spinodeflandrella* cf. *sinuata*, and one Late Permian assemblage with *Follicucullus* cf. *scholasticus*. The most representative assemblage is that with *Albaillella indensis* of the Tournaisian-Visean. Three species of this assemblage (*Albaillella paradoxa* Deflandre, *A. indensis* Won, and *A. uncus* Won) are described.

Key words: *the Carboniferous, the Permian, radiolarians, Yakutiya.*

Late Paleozoic radiolarians are widespread in different regions of the world: in Europe, North America, Japan, China, Russia, and Kazakhstan, where they are adequately studied. The radiolarians occur in deposits of different (carbonate, siliceous, and terrigenous) lithological types. The most representative and diverse Late Paleozoic radiolarians were recovered from carbonate concretions of the South Uralian flysch deposits, which yielded well-preserved and abundant forms of the Late Carboniferous and Early Permian age (Nazarov and Rudenko, 1981; Kozur, 1980, 1981; Isakova and Nazarov, 1986; Nazarov, 1988; Amon, 1985; Amon *et al.*, 1990). In addition, the radiolarian assemblage with *Polyentactinia nyatvica* referred to the Middle Carboniferous (the Moscovian Stage) was identified by Nazarov in carbonate concretions from the left bank of the Kolyma River, the Kolyma Uplift (Nazarov and Ormiston, 1987, 1993; Nazarov, 1988).

The Late Paleozoic radiolarians are widespread in siliceous deposits of Russia. Nazarov reported on the Early Carboniferous poorly preserved radiolarians *Albaillella* sp., *Pylentonema* sp., *Popofskyellum* sp., and *Palaeoscenidium* (?) sp. from the siliceous deposits of Middle Asia and eastern Kazakhstan, and on the assemblage with *Albaillella paradoxa* from jaspers and phtanites of the Penzhina Ridge, all studied only in thin sections. Regretfully, it is unknown whether Nazarov later distinguished any Carboniferous radiolarians from siliceous deposits by means of chemical treatment. He briefly mentioned just their occurrence in one of his last publications (Nazarov, 1988, p. 144). The Early Carboniferous radiolarians (*Archocyrium* sp., *Entactinia* aff. *vulgaris*, and *Pylentonema* sp.), along with some Devonian and Ordovician forms, were discovered in chert samples from the melange and olistostrome of the Chara belt in eastern Kazakhstan (Ivata *et al.*, 1994). The Late Carboniferous radiolarians, including *Tormentum sequilateralis* Nazarov, *T.* cf. *pervagatum* Nazarov, and *Latentifistula* sp., also were identified in the siliceous deposits of the Koryak Upland (Krimsalova, 1994). In Russia, Permian radiolarians are known from siliceous deposits of the Sikhote-Alin' (Rudenko and Panasenko, 1990b; Rudenko, 1991; Nikitina *et al.*, 1992; Rudenko, 1994), the Koryak Upland (Rudenko, 1991; Vishnevskaya, 1994), and Sakhalin (Bragin, pers. commun.).

Single Late Permian radiolarians were found in the terrigenous siltstones and tuffaceous siltstones of the Vladivostok, Ugodinzin and Yastrebovka formations of Primor'e (Belyanskii *et al.*, 1984; Kiseleva and Rudenko, 1984; Rudenko and Panasenko, 1990a; Rudenko, 1991).

The new occurrences of radiolarians of the Late Paleozoic are very important, because this time interval comprises the stages of thriving and the global extinction of one of the largest group of bilaterally symmetrical radiolarians of the Albaillellaria Order. The radiolarian localities of Yakutiya mark the northernmost geographic position of Albaillellaria distribution.

† Deceased.

A comparative analysis of the studied assemblages with those known from other regions can improve our understanding of the evolution and paleogeographic distribution of Late Paleozoic radiolarians. The new data are also valuable in a stratigraphic aspect, especially for dating and the subdivision of siliceous deposits.

This article is based on the analysis of the Late Paleozoic radiolarians that we detected in siliceous tuffs and tuffites sampled at the left bank of the Indigirka River in the Selennyakh block of northeastern Yakutiya (Fig. 1). The Selennyakh block extends meridionally for 120 km in the interfluve between the middle course of the Uyandina River and the upper reaches of the Selennyakh River. The block is predominantly composed of Ordovician, Silurian and Devonian carbonates and the Lower Carboniferous clayey-siliceous deposits, which are deformed into large folds and monoclines disturbed by the reversed and strike-slip faults of northeastern and meridional orientation. Tectonic structures of the Selennyakh block are constituents of the Kolyma–Omolon microcontinent (Parfenov, 1991). The Selennyakh block is separated by the submeridional Tirekhtyakh fault from the Triassic and Jurassic deposits of the southern Polousnyi synclinorium in the west.

Early Permian (Early Asselian) radiolarians have been first identified by S.V. Zyabrev (Institute of Tectonics and Geophysics, Far East Division of Russian Academy of Sciences) in 1991 from a single sample collected by A.V. Nesterenko in the southern limb of the Polousnyi synclinorium adjacent to the Selennyakh Block, in the distribution area of the Triassic and Jurassic deposits (Fig. 2). The assemblage with *Latentifistula astricta* that is detected here will be considered in detail below together with other assemblages. The radiolarian-bearing sequence is composed of virtually unaltered pelitomorphic gray and greenish gray, frequently banded siliceous tuffite and rare interbeds of fine-grained sandstone. The deluvial talus of the sequence is exposed at the left watershed of the Nagondzha River. The sequence is estimated to be not more than 100 m thick. Stratigraphic relationships between the sequence and underlying and overlying deposits cannot be observed, but, according to the geological survey data, the sequence rests on fossiliferous limestones and conglomerates of the Lower Devonian (Emsian Stage) and underlies the terrigenous-carbonate deposits containing bivalves of the Upper Triassic (Norian Stage). In the general structure, it is located in the core part of the anticline (Fig. 2). In 1993–1994, we collected some samples of the siliceous tuffites from the sequence. The samples yielded a rich radiolarian assemblage with *Albaillella indensis* of the Early Carboniferous age (Rudenko *et al.*, 1994). Thus, in the area under consideration, the Early Carboniferous and Early Permian radiolarians were first distinguished from the siliceous tuffite sequence of inconsiderable thickness. In 1994, V.A. Aristov identified some Early Carboniferous conodonts from the same sequence.

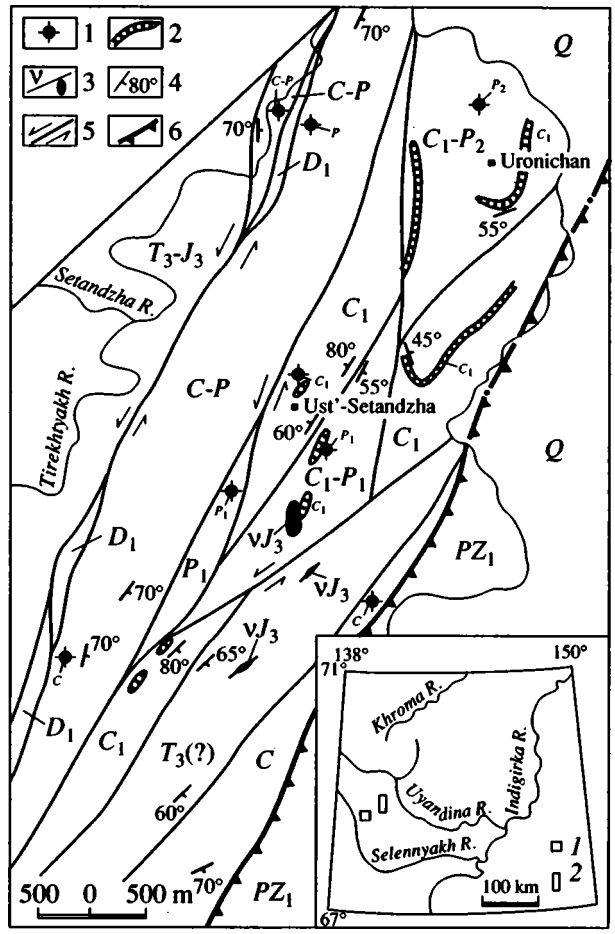


Fig. 1. Geological structure of the area in the right bank of the Tirekhtyakh River: (1) radiolarian localities; (2) marker limestone beds; (3) dikes and bodies of the Upper Jurassic gabbro-diabase; (4) attitude and dip of beds; (5) normal and reverse strike-slip faults; (6) tectonic boundary separating Lower Paleozoic carbonate deposits from younger formations. The inset map shows the Late Paleozoic radiolarian localities: (*1*) in the Nagondzha River basin; (*2*) in the Tirekhtyakh River basin.

An area of the Lower Carboniferous–Permian deposits up to 2 km wide and up to 12 km long stretches northeastward along the right bank of the Tirekhtyakh River (the right tributary of the Uyandina River) as a constituent of the northern Selennyakh block and is located 50 km to the north of the described region. These deposits are faulted by minor shears and contact along the reversed strike-slip faults with clayey-siliceous deposits of Late Triassic–Early Jurassic age in the west and with sandstones and siltstones of presumably Late Triassic age and carbonate Lower Paleozoic deposits in the east (Fig. 1). It is impossible to observe stratigraphic relationships of these deposits with underlying and overlying rocks.

Some fragments of this sequence were studied in scarce bedrock exposures and in alluvial-deluvial deposits. The sequence is represented by alternating

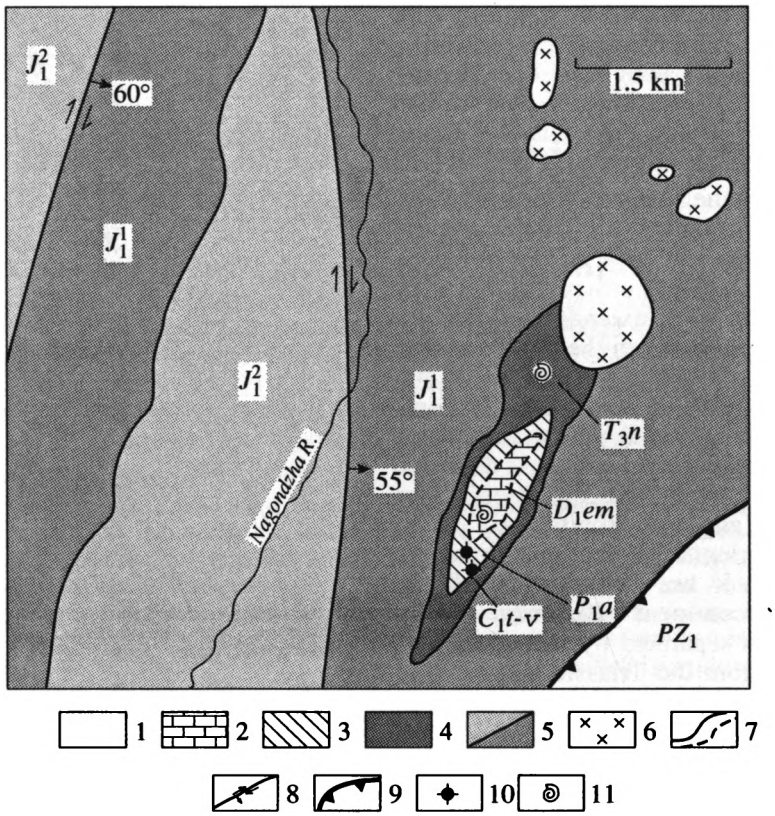


Fig. 2. Geological structure of the area in the left bank of the Nagondzha River: (1) Middle Paleozoic carbonate deposits; (2) Devonian carbonate deposits; (3) Carboniferous-Permian siliceous tuffites; (4) Triassic carbonate-terrigenous deposits; (5) Jurassic terrigenous deposits; (6) diorite; (7) established and presumable stratigraphic boundaries; (8) strike-slip fault separation plane dip; (9) overthrust; (10) localities of radiolarians; (11) localities of macrofauna. The geographic position of the area is shown in the inset map of Fig. 1.

fine-grained to pelitomorphic siliceous tuffites and tuffaceous shales of dark-gray, black, and greenish gray color. The deposits include beds and lenses of dark- and light-gray fine-grained massive limestones and calcareous siltstones. The siliceous tuffites have a pelitic texture and shaly to massive structure. The fine-grained pelitic clayey mass contains from 5 to 30% of micrograined siliceous matter. It also bears abundant radiolarian tests filled with secondary material. The limestones are composed of newly formed carbonate grains isometric in form and variable in size (0.3–0.8 mm). Prospect trenches revealed stratigraphic contacts of tuffites and pelitic tuffs with limestones. The thickness of limestone beds varies from 0.5–1 to 15–20 m; the thickest beds traceable for 300–1300 m in the watershed area and easily recognizable in aerial photographs are used as the marker horizons. By geological mapping on a scale of 1 : 200 000, A.V. Dorofeev (unpublished data of 1969) discovered limestone beds with foraminifers of the Early Carboniferous (Tournaisian-Visean). Afterwards, special investigations conducted by A.V. Nesterenko (1992) detected some silicified columnals of crinoids in a limestone bed near the town of Uronichan, and G.A. Stukalina (All-Russia Geological Institute) assigned them to the Visean and Namurian stages of the Lower Carboniferous. In 1990, V.P. Tarabukin distinguished some Early Carboniferous (Tournaisian) conodonts from a limestone bed exposed further to the south. Carboniferous and Permian radiolarians were found in limestones, siliceous tuffites, and pelitic tuffs (Fig. 1). In 1993, N.Yu. Bragin established three assemblages of radiolarians from these deposits: the Early Carboniferous assemblage with *Albaillella* cf. *paradoxa* Deflandre, the Early Permian assemblage with *Spinodeflandrella* cf. *sinuata*, and the Late Permian assemblage with *Follicucullus* cf. *scholasticus* Ormiston et Babcock.

Thus, the hemipelagic deposits of northeastern Yakutia contain the following radiolarian faunas: the Early Carboniferous assemblages with (1) *Albaillella* cf. *paradoxa* and (2) *Albaillella indensis*; the Early Permian age assemblages with (3) *Latentifistula astricta* and (4) *Spinodeflandrella* cf. *sinuata*; and the Late Permian assemblage (5) with *Follicucullus* cf. *scholasticus*. Easily detectable bilaterally symmetrical and stauraxonic radiolarians of these assemblages enable the accurate dating of the enclosing deposits and interregional correlation, whereas the spherical Spumellaria

from siliceous deposits are seldom diagnostic and have wider stratigraphic ranges.

(1) The assemblage with *Albaillella* cf. *paradoxa* is extracted from the siliceous tuffs of the Tirekhtyakh River basin. It consists of *Albaillella* cf. *paradoxa* Deflandre, *A.* cf. *uncus* Won, *Archocyrtium* sp., *Cyrtisphaeroctenium* (?) sp., Entactinia cf. *vulgaris* Won, and *Spongoentactinia* cf. *variabilis* (Ormiston et Babcock). The species *Albaillella paradoxa* is widespread in the following localities of the Lower Carboniferous deposits: the Tournaisian phosphate concretions of the Baltalimani Formation, the Bosporus, Turkey (Holdsworth, 1973); the Tournaisian phosphate concretions of the Montagne Noire (Deflandre, 1952) and Central Pyrenees, France (Gourmelon, 1986, 1987); the Tournaisian calcite concretions in the phosphate beds of the Woodman Formation, Utah, United States (Gourmelon, 1987; Sandberg and Gutschick, 1984); and the middle Tournaisian of siliceous shales and mudstones of "Kulm-facies", Germany (Braun and Schmidt-Effing, 1993). The assemblage with *Albaillella* cf. *paradoxa* is certainly the Early Carboniferous (most probably Tournaisian) in age.

(2) The assemblage with *Albaillella indensis* is most significant and diverse. These radiolarians were found in siliceous tuffites of the Nagondzha River basin. The assemblage is mainly represented by spherical forms of a poor preservation complicating identification of genera and species. Among them, there are *Entactinia vulgaris* Won, *Entactinia* cf. *vulgaris* Won, *E.* cf. *parva* Won, *E.* (?) *inaequoporosa* Won, *Entactinia* sp., *Meschedea hirsuta* Won, *M.* cf. *pyramispinosa* Won, *Belowea tenuestesta* Won, *Belowea* cf. *variabilis* (Ormiston et Lane), *Belowea* sp., and *Triaenosphaera* cf. *bareillensis* Gourmelon. Another abundant group includes representatives of the *Albaillella* genus: predominant *Albaillella indensis* Won (hundreds of specimens) and rare *A. uncus* Won, *A.* cf. *paradoxa* Deflandre, and *Albaillella* sp. Both groups coexist with infrequent *Popofskyellum* cf. *campanella* Won, *Popofskyellum* sp., *Archocyrtium* cf. *coronaesimile* Won, *A.* cf. *riedeli* Deflandre, *A.* cf. *lagabirielli* Deflandre, *Archocyrtium* sp., *Pylentonema* sp., *Ceratoikiscum* sp., *Cubaxonium* (?) *octaedrospongiosum* Won, and *Palaeoxyphostylus* (?) sp.

The assemblage with *Albaillella indensis* is widespread in the Lower Carboniferous of Europe, North America, and Asia. It is most similar with the assemblage from the "Gedauer-Konglomerat" of Germany (Won, 1983). The *Albaillella indensis* Zone was also recognized in the phosphate concretions of the Montagne Noire in France (Gourmelon, 1987), where its stratigraphic range corresponds to the upper Tournaisian–lower Visean. The index species is also known in the "faunal Group IA" of the northern Brooks Range (Holdsworth and Murchey, 1988). A similar assemblage was distinguished as well in siliceous deposits of China, in the Shixia Reservoir of southeastern Guangxi (Wang and Kuang, 1993). In addition, the *Albaillella indensis* Zone was recognized in Germany (Braun and Schmidt-Effing, 1993). It is referred here to the upper Tournaisian–lower Visean. Thus, we consider the *Albaillella indensis* assemblage from the siliceous tuffs of the Yakutiya to be of the Tournaisian–Visean age

(3) The *Latentifistula astricta* assemblage (according to data of Zyabrev) from the siliceous tuffites of the Nagondzha River basin is mainly represented by stauraxonic radiolarians *Latentifistula astricta* Nazarov, *L. astricta astricta* Nazarov, *L. torulosa* Nazarov, *Latentifistula* aff. *crux* Nazarov et Ormiston, *Tormentum* aff. *circumfluum* Nazarov, *Tormentum* spp., *Copicyntra* spp., and *Spongentactinia* (?) spp. Similar radiolarian assemblages are known from the Lower Permian of the southern Urals and North America (Nazarov and Ormiston, 1985).

(4) The *Spinodeflandrella* cf. *sinuata* assemblage was distinguished in limestones of the Tirekhtyakh River basin. The sparse radiolarians are represented here by *Spinodeflandrella* cf. *sinuata* (Ishiga et Watase), *Spinodeflandrella* sp., and *Latentifistula* (?) sp. The index species is widespread in the Lower Permian siliceous deposits throughout the world, being known in Japan (Ishiga, 1986, 1990), Sikhote Alin' (Rudenko and Panasenko, 1990b; Rudenko, 1991), the Koryak Upland (Rudenko, 1991; Vishnevskaya, 1994), and North America (Holdsworth and Murchey, 1988). The assemblage is similar in composition to radiolarians from the *Spinodeflandrella acutata* Zone of Sikhote-Alin' (Yakhtashian–Bolorian of the Tethian scale), and *Albaillella sinuata* Zone of Japan (Leonardian of the North American scale). Until recent times, assemblages of this type were known only from siliceous deposits. The occurrence of their radiolarians in the carbonate deposits of the Yakutia suggests a considerably wider spectrum of habitats of the Early Permian Albaillellaria.

(5) The *Follicucullus* cf. *scholasticus* assemblage was distinguished from siliceous tuffites of the Tirekhtyakh River basin. Rare radiolarians are represented here by *Follicucullus* cf. *scholasticus* Ormiston et Babcock and *Tormentum* cf. *inflatum* Nazarov et Ormiston. *Follicucullus scholasticus* is widespread in carbonate, terrigenous and siliceous deposits of the Upper Permian in Japan (Ishiga, 1986, 1990; Caridroit, 1986; and others), the Sikhote–Alin' (Belyanskii *et al.*, 1984; Rudenko and Panasenko, 1990a), and North America (Ormiston and Lane, 1979). The *Follicucullus scholasticus* assemblage from Sikhote–Alin' can be assigned to the Murgabian Stage of the Tethyan scale, or to the upper Guadelupian Stage of the North American scale.

Thus, the Late Paleozoic radiolarians were first found in Yakutiya, and the Early to Late Permian radiolarians enabled us to date the enclosing deposits of the region, whereas the Early Carboniferous radiolarians

from siliceous deposits are distinguished and described in Russia for the first time.

The Late Paleozoic radiolarians occur in hemipelagic deposits reduced in thickness. Analogous siliceous and siliceous-clayey sequences with thin limestone interbeds are known from the western framing structures of the Selennyakh block and from the Tas-Khayakhtakh block of the Kolyma–Omolon microcontinent located to the south. In many cases, these sequences are considered free of fossils, and their age is conventionally dated as the Early Carboniferous or Early Jurassic. Similar sequences in other orogenic belts do not exceed several dozens or few hundreds of meters in thickness. Their accurate dating requires a thorough study of microfauna, primarily, radiolarians.

The data obtained may help to reconstruct the Late Paleozoic geological history of the region. The northeastern Yakutiya experienced rifting in the second half of the Paleozoic Era. At that time, the Selennyakh block was a component of the Omulevskii microcontinent separated by oceanic basins from the Verkhoyansk passive margin in the west and from the Kolyma–Omolon microcontinent in the east (Parfenov, 1991). The Late Paleozoic deposits of the Verkhoyansk passive margin were accumulated in different facies environments as compared to the described sequences deposited in an open sea, far from the Kolyma–Omolon microcontinent and this passive margin. Undoubtedly, the oceanic basin was much greater than the distribution area of the exposed sequences. The current structural position of these sequences seems to be a result of the Late Mesozoic collision between the North Asian craton and the Kolyma–Omolon microcontinent that was responsible for subsequent horizontal displacements and resultant spatial proximity of different sedimentary facies of the Late Paleozoic age.

Because the Carboniferous radiolarians are found sporadically in the Russian territory and their new locality is confined to siliceous deposits located far from the type area, it seems necessary to describe once more the most significant species of Early Carboniferous *Albaillella* genus from Yakutiya.

Albaillella indensis Won, 1983,
Plate, Figs. 1–4

1983. *Albaillella indensis* Won, p. 127; Pl. 1, Figs. 19 and 20; Text-Fig. 3g.

1987. *Albaillella indensis* Won: Gourmelon, pp. 87–88, Pl. 12, Figs. 6–10.

1988. *Albaillella indensis* Won group: Holdsworth and Murchey, p. 789, Pl. 34.1, Figs. 16–20.

1993. *Albaillella indensis* Won: Braun and Schmidt-Effing, Pl. II, Fig. 4.

1993. *Albaillella indensis* Won: Wang Yu-jing and Kuang Guo-dun, Pl. I, Figs. 1–8.

Description. Bilaterally symmetrical cone-shaped cross-segmented (5–7 segments) test has a spinelike process in the lower part of the apical cone.

Size (based on 32 specimens): the test without H-frame is 200–250 μm in height and 120–150 μm in diameter.

Comparison. The species differs from *Albaillella paradoxa* Deflandre (Deflandre, 1952, pp. 872–874, Figs. 1–3, 5) in having the segmented test, spinelike process, and larger diameter of the lower part of the test. The studied specimens also differ by smaller tests from the earlier described representatives of this taxon.

Locality: siliceous tuffites at the left bank of the Indigirka River (the Nagondzha River basin), northeastern Yakutiya.

Distribution: the Tournaisian of France, phosphate concretions of the Central Pyrenees (Gourmelon, 1986, 1987); the Namurian (?) of Germany, "Gedauer-Konglomerat" (Won, 1983); the Tournaisian (?)–Visean (Middle–Upper Osagen, Lower Mississippian) of North America, siliceous deposits of the northern Brooks Range (Holdsworth and Murchey, 1988); the Tournaisian–Visean of Germany, siliceous shales and mudstones of "Kulm-facies" (Braun and Schmidt-Effing, 1993); the Tournaisian–Visean of China, siliceous deposits of southeastern Guangxi (Wang and Kuang, 1993).

Albaillella paradoxa Deflandre, 1952

Plate, Fig. 5

1952. *Albaillella paradoxa* Deflandre, pp. 872–874, Figs. 1–3, 5.

1966. *Albaillella paradoxa* Deflandre: Holdsworth, pp. 321–323.

1969. *Albaillella paradoxa* Deflandre: Holdsworth, pp. 230–236, Pl. I, Figs. 10a and 12a.

1973. *Albaillella paradoxa* Deflandre: Holdsworth, p. 127, Pl. I, Figs. 12 and 15.

1985. *Albaillella paradoxa* Deflandre: Gourmelon, Pl. 2, Fig. 20.

1986b. *Albaillella paradoxa* Deflandre: Gourmelon, pp. 191–192, Pl. 3, Fig. 5; Pl. 4, Fig. 4.

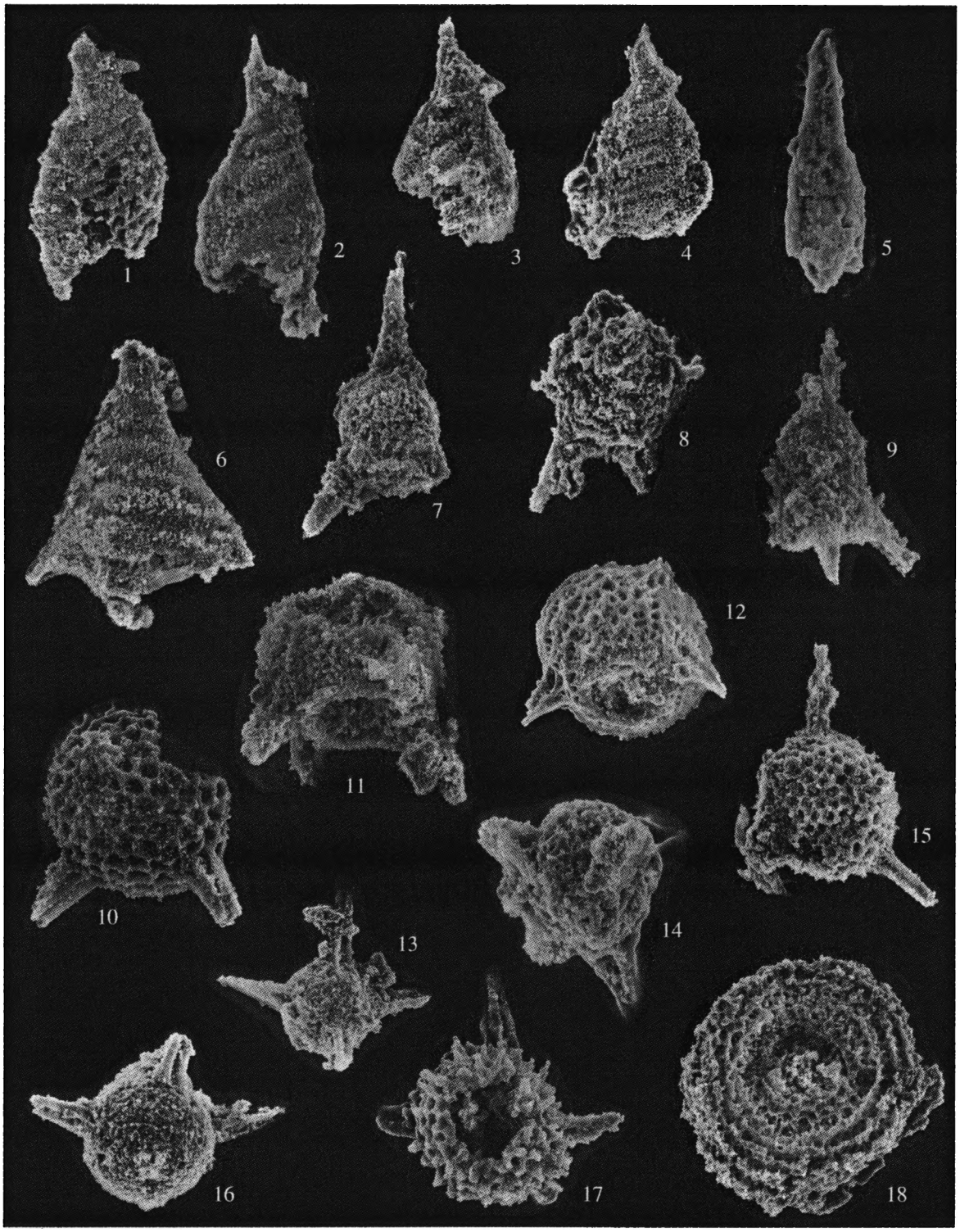
1987. *Albaillella paradoxa* Deflandre: Gourmelon, pp. 84–85, Pl. 11, Figs. 1–5.

1993. *Albaillella paradoxa* Deflandre: Braun and Schmidt-Effing, Pl. 1, Fig. 6.

Description. The test is bilaterally symmetrical, conical, straight and smooth.

Size (based on 3 specimens): the test without H-frame is 240 μm in height and 80 μm in diameter.

Comparison. This species differs from *Albaillella cornuta* Deflandre (Deflandre, 1952, p. 873, Figs. 6 and 7) in the absence of a spinelike process, and, in contrast to other species of the genus, it has the straight conical test.



Early Carboniferous (Tournaisian–Visean) radiolarians from siliceous tuffites of the left bank of the Indigirka River, northeastern Yakutiya

Assemblage with *Albaillella indensis:* (1–4) *Albaillella indensis* Won, ×160; (5) *Albaillella paradoxa* Deflandre, ×160; (6) *Albaillella uncus* Won, ×250; (7) *Archocyrtium* cf. *lagabriellei* Deflandre, ×160; (8) *Archocyrtium* sp., ×250; (9) *Archocyrtium* cf. *riedeli* Deflandre, ×250; (10) Sphaeroidea, gen. et sp. indet., ×160; (11) *Pylentonema* sp., ×250; (12) *Pylentonema* sp., ×160; (13) *Belowea* (?) sp., ×160; (14) *Triaenosphaera* (?) sp., ×250; (15) *Triaenosphaera* cf. *bareillensis* Gourmelon, ×160; (16) *Pylentonema* sp., ×160; (17) *Entactinia* cf. *inaequoporosa* Won, ×160; (18) *Belowea variabilis* (Ormiston et Lane), ×160.

Locality: siliceous tuffs and tuffites at the left bank of the Indigirka River (basins of the Nagondzha and Tirekhtyakh rivers), northeastern Yakutiya.

Distribution: the Tournaisian of Turkey, phosphate concretions of the Baltalimani Formation, the Bosphorus (Holdsworth, 1973); the Tournaisian of France, phosphate concretions of the Montagne Noire (Deflandre, 1952) and Central Pyrenees (Gourmelon, 1986, 1987); the Tournaisian of North America, calcitic concretions in phosphate beds of the Woodman Formation, Utah, United States (Sandberg and Gutschich, 1984 after Gourmelon, 1987).

Albaillella uncus Won, 1983,
Plate, Fig. 6

1983. *Albaillella uncus* Won, p. 127, Pl. 1, Figs. 16–18, Text-Figs. 3d and 3e.

Description. The test is conelike, segmented (7 segments), and has two wings in the lower part.

Size (based on 2 specimens): the test without H-frame is 190 μm in height and 150 μm in diameter at the wing base.

Comparison. The species differs from morphologically similar *Albaillella furcata* Won (Won, 1983, pp. 126–127, Pl. 12, Figs. 3–5, 7) in smaller test size and less developed wings.

Locality: siliceous tuffs and tuffites at the left bank of the Indigirka River (the basins of the Nagondzha and Tirekhtyakh rivers), northeastern Yakutiya.

Distribution: the Namurian (?) of Germany, "Gedauer–Konglomerat" (Won, 1983).

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BRIEF COMMUNICATIONS

The Age of Volcanic Sequence in the Middle Paleozoic Deposits of Southern Mongolia

V. A. Aristov* and G. Nyamsuren**

** Geological Institute (GIN), Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia*
*** Geological Institute, Mongolian Academy of Sciences, 210351 Ulan Bator, Mongolia*

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Abstract—Recent finds of conodonts indicate that the volcanogenic sequence in the Gobi Altai–Suhe Bator Zone of southern Mongolia (the Gobi Altai Formation in the Dzhinsetu Ulul Range area and the Minzhin Formation in the Mandal Obo area), earlier regarded as barren of fossils, was formed in the interval between the Givetian of the Middle Devonian and the early Fammennian of the Late Devonian. The ratio of effusive and volcaniclastic rocks in the studied sections suggests that centers of volcanism moved from east to west during the indicated period.

Key words: *Stratigraphy, Devonian, conodonts, Mongolia.*

Volcanic rocks play an important role in the Devonian sections of southern Mongolia. They form thick (up to 2 km) beds or separate members among terrigenous–carbonate deposits, being represented by various tuffs, tuffstones, tuffites, and siliciliths intercalated with lava flows of a widely ranging composition, from rhyolites to basalts. The distribution area of these rocks has a complicated block and fold–nappe structure.

This article is devoted to the volcanogenic sequence progressively thickening the terrigenous–carbonate complex of the Lower–Middle Devonian (Eifelian) deposits accumulated in the South Mongolian Variscan paleobasin (the southern shelf of the North Asian Caledonian continent). The rock complex extends in a form of a narrow sublatitudinal strip throughout southern Mongolia for a distance of more than 1600 km, and, in recent structures, it is confined to the Gobi Altai–Suhe Bator tectonic zone (Ruzhentsev and Pospelov, 1992).

Volcanic formations are either completely devoid or contain few remains of benthic macrofauna, traditionally used for stratigraphic subdivision of the Paleozoic. Therefore, the age of volcanogenic strata was for a long time poorly defined, as the dating results were not precise. They spanned a considerable interval, and sometimes were conventionally based on the bed position in the section and on fauna finds in the underlying and overlying beds, or in the neighbouring blocks.

The geology of southern Mongolia has been studied in detail during the last two decades by the Joint Soviet–Mongolian Geological Expedition. In 1989–1990, we used bed-by-bed sampling for the conodont analysis in order to solve problems of the Devonian–Lower Carboniferous stratigraphy. It turned out that these microscopic organic remains, formerly unstudied in the region, are very useful for the subdivision of Paleozoic marine strata in southern Mongolia, including the volcanogenic members.

Stratigraphic studies of volcanogenic formations were undertaken in the Gobi Altai mountains at two sections situated at a distance of 400 km from each other (Fig. 1). One is on the southern slope of the Gobi Altai (the Dzhinsetu Ulul Range near the Shine Dzhinst Somon), and another is farther east (the Gobi Altai northeastern hills near the Mandal Obo Somon). Below, we shall separately review the structure of the volcanogenic sequence in two sections.

DZHINSETU ULUL RANGE

The volcanogenic sequence of this locality was studied on the right bank of the Bayan Sair Creek, 2 km east of the Tsahirin Khuduk, where the Emsian–Eifelian reference section is rich in fossils (Alekseeva *et al.*,

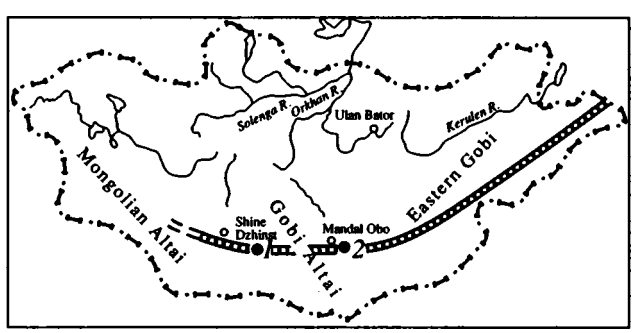


Fig. 1. Geographic scheme of studied localities: (*1*) Dzhinsetu Ulul, (2) Mandal Obo; the hatched area indicate the Gobi Altai–Suhe Bator tectonic zone.

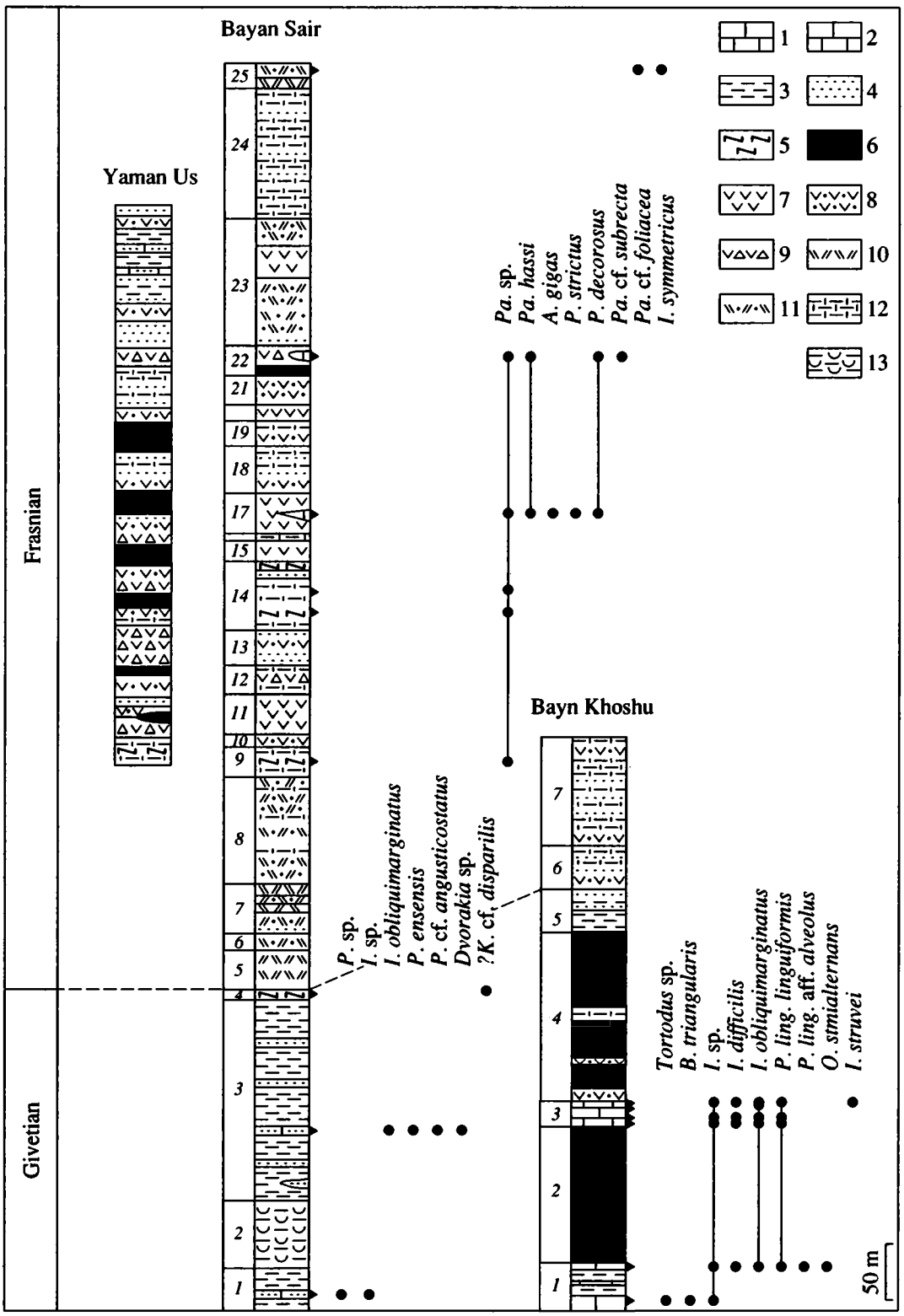


Fig. 2. Correlation of volcanic sections and distribution of conodonts: (*A.*) *Ancyrodella*, (*B.*) *Belodella*, (*I.*) *Icriodus*, (*O.*) *Ozarkodina*, (*P.*) *Polygnathus*, (*Pa.*) *Palmatolepis*; (1) crystalline limestone, (2) sandy limestone, (3) siltstone and shale, (4) sandstone and tuffstone, (5) phtanite, (6) basalt, (7) plagioclase porphyry, (8) basic tuff, (9) tuff-breccia, (10) quartz–plagioclase porphyry, (11) tuffs of the intermediate and acid composition, (12) siliceous tuffites and tuffaceous siltstone, (13) ashy tuffite; bed numbers are shown in the left columns.

1996). The volcanic–terrigenous rocks of this region have been recently united into the Gobi Altai Formation (Alekseeva, 1993).

Above the crynoidal limestones with diverse Eifelian organic remains, the Bayan Sair section consists of the following successively deposited rock units (Fig. 2):

The Gobi Altai Formation

(1) Siltstone, dark gray, schistose. The lower part of the bed contains a thin layer of sandy limestone with tentaculitids, fish remains, and conodonts *Polygnathus* sp., *Icriodus* sp., (36.5 m).

(2) Ashy tuffite with obscure foliation, light green (60 m).

(3) Siltstone, greenish gray; in the upper part, this bed is black, foliated, and displaying interlayers and lenses of slightly calcareous sandstone up to 1m in thickness. An interbed of sandy limestone, identified 58 m above the basal part, yielded small gastropods, tentaculitids, and conodonts *Polygnatus ensensis* Ziegl. et Klapp., *P.* cf. *angusticostatus* Witt., *Icriodus obliquimarginatus* Bisch. et Ziegl., and *Dvorakia* sp. (180 m).

(4) Phthanite, laminated and bearing conodonts: ?*Klapperina* cf. *disparilis* (Ziegl., Klapp. et Johns.) (3.2 m).

(5) Quartz–plagioclase porphyry (40 m).

(6) Tuff of the intermediate composition, crystalloclastic (15 m).

(7) Tuffstone, green (3 m), and alternating coarse-grained tuffs and porphyries (45 m) higher in the section.

(8) Tuff, fine- to medium-grained with interlayers of siliceous tuffs (100 m).

(9) Tuffite, siliceous, with interlayers of silty and radiolarian phthanites, the latter bearing *Palmatolepis* sp. (25 m).

(10) Tuff, crystalloclastic, basic in composition and comprising interlayers of fine-grained tuffites (12 m).

(11) Plagioclase porphyry, dark green (36 m).

(12) Tuffite, siliceous, green, comprising interlayers of tuffaceous sltstones and a horizon of tuffbreccias in the upper part (28 m).

(13) Tuffstone, tobacco green, with an interlayer of tuff of the basic composition (30 m).

(14) Tuffite, siliceous, green, with scarce phthanite interlayers. The upper part of the bed comprises an interlayer of tuffstone and sill of plagioclase porphyrite. Conodonts of the unit are represented by *Palmatolepis* sp. (62 m).

(15) Plagioclase porphyrite (17 m).

(16) Tuffite, siliceous, with fine rhythmic bedding (6 m).

(17) Plagioclase porphyrite, green, 16 m above the base enclosing a thin lens of pink limestone with crinoids, ostracods, and conodonts *Ancyrodella gigas* Youngq., *Palmatolepis hassi* Mill. et Mill., *P.* sp., *Polygnathus decorosus* Stauf., *P. strictus* Kuzmin et Yurtch (40 m).

(18) Tuff, lithocrystalloclastic, grading into sand and silt (40 m).

(19) Tuffaceous siltstone with an interlayer of crystalloclastic tuff in the lower part (25 m).

(20) Plagioclase porphyrite (13 m).

(21) Tuff of the basic composition (30 m).

(22) Basalt, green (5 m), grading upward into tuff lapilli of the basic composition, and then into tuff conglobreccia with blocks of recrystallized limestones up to 0.8 m across. One of the blocks contains conodonts *Palmatolepis hassi* Mill. et Mill., *P.* cf. *subrecta* Mill. et Youngq., *P.* sp., *Polygnathus decorosus* Stauf., and *Belodella* sp. (26 m).

(23) Tuffstone locally enclosing interlayers of lithocrystalloclastic tuff. A porphyrite body occurs 60–90 m above the bed base below crystalloclastic tuff (115 m).

(24) Tuffite, siliceous, banded, containing interlayers of tuffstones at the level of 40 m above the base and higher (120 m).

(25) Quartz–plagioclase porphyry grading into, crystalloclastic coarse-grained tuff. At the level of 10 m, the bed includes a lens of marmorized limestone (4.5 m) with conodonts *Palmatolepis* cf. *foliacea* Youngq., *Icriodus symmetricus* Br. et Mehl., and *Polygnathus* sp. (17 m).

After a nonexposed interval extending for 110 m, the tectonic block displays a bed of siliceous tuffite and tuffstone, 250 m thick, duplicating the upper part of the studied section. Carboniferous terrigenous deposits are recorded north of the fault.

The apparent thickness of the Gobi Altai Formation in the Bayan Sair section is 1100 m.

In the mentined section, conodonts were identified at nine levels. The conodont assemblages of three lower levels (beds 1–4) are characteristic of the Givetian Stage of the Middle Devonian, whereas those of five upper levels (beds 5–25) are typical of the Frasnian of the Upper Devonian. The rare occurrence of conodonts in the section and their low taxonomic diversity makes it difficult to distinguish conodont zones in the studied deposits, and leave the stratigraphic subdivision at the stage rank.

In addition, the early Famennian conodonts (the sample for analysis of microfaunal was collected by S.V. Ruzhentsev) were identified in an interbed of jasperoid chert intercalated with tuffites in the uppermost part of the Gobi Altai Formation at the left bank of the Bayan Sair Creek. Their assemblage includes *Palmatolepis perlobata schindewolfi* Mull., *P.* ex gr. *delicatula* Br. et Mehl, and *Polygnathus* sp; it is important for defining the upper age limit of the Gobi Altai Formation, as the latter has normal contacts with the overlying upper Famennian–Tournaisian carbonate deposits at the left bank of the Bayan Sair Creek.

Hence, the age of the Gobi Altai Formation in the Dzhinsety Ulul Range spans the interval from the

Givetian of the Middle Devonian to early Famennian of the Late Devonian.

The Givetian and Upper Devonian parts of the section considerably differ in lithology. The Givetian deposits are mostly represented by terrigenous rocks (dark-colored silty pelites with subordinate interbeds of sandstones), and only in the lower part, there is one horizon of ash tuffite 60 m thick. The Frasnian–lower Famennian part of the section is mainly composed of volcanic and volcanoclastic rocks.

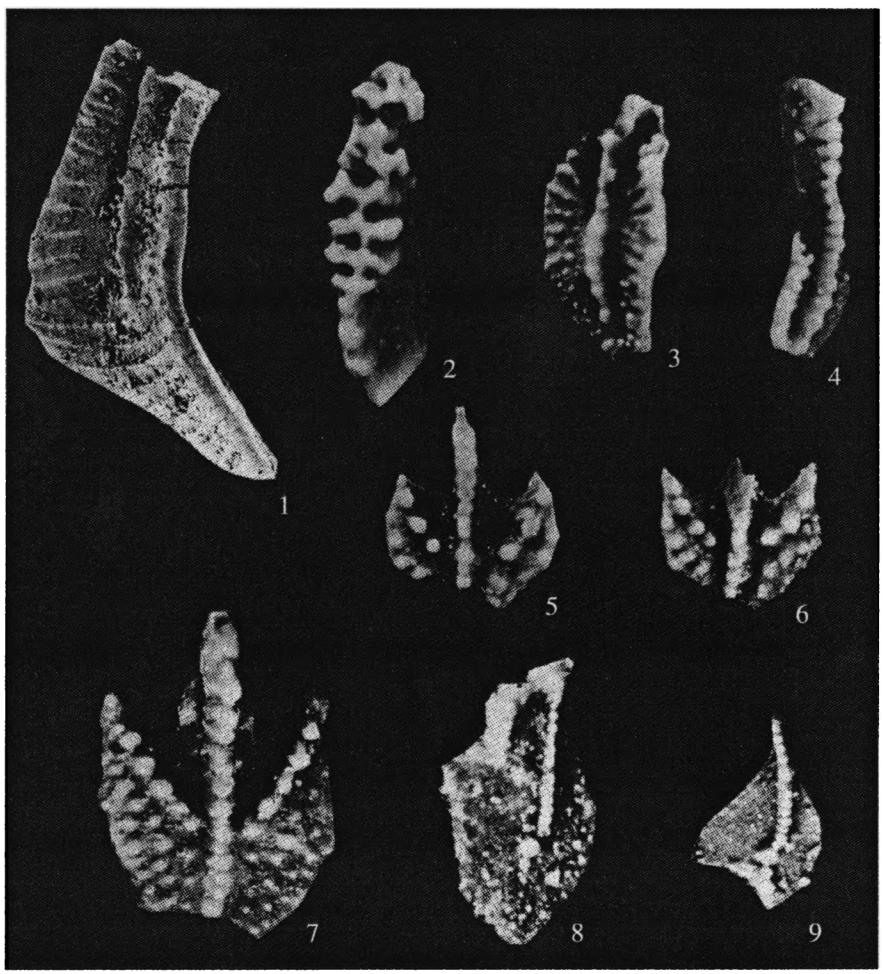
The Fransnian part of the Gobi Altai Formation was also studied in another section situated 4 km west of the Bayan Sair section, near the Yaman Us source, where volcanic units, 500 m thick, are exposed in a tectonic block (Fig. 2, left column). While the Frasnian part of the Bayan Sair section includes only one thin body of basalt and several sill-like horizons of plagioclase porphyries, the lower half of the Yaman Us section shows basaltic rocks at six levels, where they occur together with coarse lithoclastic tuffs, including lapilli, and bombs. The thickness of basaltic bodies varies from few to 25 m.

MANDAL OBO LOCALITY

In this region, volcanogenic rocks are members of the Minzhin Formation (Alekseeva, 1993), whose lower contact is entirely tectonic.

Half a kilometer east of the Bayan Khoshu ruins, the structure of the Minzhin Formation is as follows (Fig. 2, right column):

(1) Limestone, thin-bedded, organoclastic with brachiopods and tentaculitids, enclosing interbeds of greenish gary siltstone. The limestone contains conodonts *Icriodus difficilis* Ziegl., Klapp. et Johns., *I. obliquimarginatus* Bisch. et Ziegl., *I.* sp., *Polygnathus lin-*



Conodonts from the Gobi Altai and Minzhinovsk formations (southern Mongolia)

(1) *Polygnathus linguiformis linguiformis* Hinde, the Minzhinovsk Formation, Bayan Khoshu section, Bed 1; (2) *Icriodus obliquimarginatus* Bisch. et Ziegl., the Gobi Altai Formation, Bayan Sair section, Bed 1; (3) *Polygnathus strictus* Kuzm. et yurtch., ditto, Bed 17; (4) *Polygnathus decorosus* Stauf., ditto, Bed 22; (5–7) *Ancyrodella gigas* Youngq., ditto, Bed 17; (8) *Palmatolepis* sp. (*P.* cf. *hassi* Mill. et Mill.), ditto, Bed 17; (9) *Palmatolepis hassi* Mill. et Mill., ditto, Bed 22; magnification 40 for all conodont images.

guiformis linguiformis Hinde, *P. linguiformis* aff. *alveolus* Wedd., *Ozarkodina semialternans* (Wirth), *Beladella triangularis* (Stauf.), and *Tortodus* sp. (2.8 m).

(2) Basalt, amygdaloid, green (120 m).

(3) Limestone variably bedded and bearing crinoids, tentaculitids, and conodonts *Icriodus obliquimarginatus* Bisch. et Ziegl., *I. struvei* Wedd., *I.* sp., *Polygnathus linguiformis linguiformis* Hinde, and *Belodella* sp. (21.6 m).

(4) Basalt, dark green, with nests of chemigenic red jasper and rare interbeds of tuff of the basic composition, and also of micrograined tuffite (155 m).

(5) Silstone, greenish gray, with interbeds of tuffstone and red jasper in the upper part (40 m).

(6) Tuffstone and red tuff in the lower part and tuffaceous siltstones with tuffstone interbeds upsection (40 m).

(7) Vaguely rhythmic tuffite (in each rhythm, several m thick, tuffstone grades into tuffaceous siltstone), variegated, with occasional interbeds of crystalloclastic tuff of the basic composition (apparent thickness about 100 m).

Higher, the volcanic sequence is overthrust by limestone beds, which yielded Tournaisian conodonts.

The thickness of the Minzhin Formation in the Bayan Khoshu section is about 500 m. Conodonts are identified only in its lower part, and, apparently, they are Givetian in age. The upper tuffite part of the section (beds 6–7) is fossil-free, but due to its conformable position on the Givetian rocks, it may be attributed to the Frasnian Stage.

Two and a half km westward from the Bayan Khoshu section, on the right bank of a dry creek bordering the Murgutsug Khuduk well, the thickness of the Minzhin Formation is 930 m (Alekseeva, 1993). Its lower part (445 m) is composed of basalts, with occasional siltstone and tuffite interbeds, and with lenses of red jasper and limestone. The upper part is essentially tuffite in composition. Here, the Minzhin Formation encloses a horizon of calcareous sandstone at the base, and is conformably overlain by limestones of the Orynshandy Formation of the Late Famennian–Tournaisian age (Alekseeva, 1993). The lowermost level with conodonts has been detected 18.2 m above the base of the formation, which contains an assemblage characteristic of the upper Famennian *Siphonodella praesulcata* Zone.

The Givetian conodonts from the basal part of the formation and late Famennian conodonts from the lower part of the overlying Orynshandy Formation show that the volcanic Minzhin Formation of the Mandal Obo locality spans an interval between the Givetian and early Famennian.

CONCLUSION

Conodonts recently found in volcanogenic sequences of the Gobi Altai–Sukhe Bator Zone of southern Mongolia (the Gobi Altai Formation of the Dzhinsetu Ula Range and the Minzhin Formation in the Mandal Obo area), suggest that volcanic rocks were formed in the interval between the Givetian of the Middle Devonian and early Famennian of the Late Devonian. The ratio of effusive to volcanoclastic rocks in the investigated sections enables us to conclude that volcanic centers migrated from east to west during the mentioned period of time.

Volcanogenic formations are more widespread in Devonian sections of the Mongolian Altai (western Mongolia), where they are concentrated at three stratigraphic levels: Lower Devonian, Lower–Middle Devonian, and Middle–Upper Devonian (Dergunov *et al.*, 1980). The studied volcanic formations can be correlated with the latter.

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BRIEF COMMUNICATIONS

The Oligocene–Miocene Boundary in the Reference Section of the Black Sea Region (Southern Ukraine)

M. F. Nosovskii

Research Institute of Geology, Dnepropetrovsk University, pr. Karla Marksa 36, Dnepropetrovsk, 320044 Ukraine

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At the 8th Joint Session of the Permanent Stratigraphic Commissions on Paleogene and Neogene of the Interdepartmental Stratigraphic Commission (ISC), a section of the Oligocene in the Black Sea region the southern Ukraine was adopted as the stratotype for the Oligocene subdivision in the territory of the former USSR (*Postanovleniya MSK...*, 1968, p. 43). Concurrently, the indicated section, the lower part of which is exposed in the Al'ma depression on the southern slope of the Mt. Kyzyl-Dzhar and which was reached by Well 1 specially drilled in the Kherson area, was endorsed as a reference section for this region. The resolution further emphasized that the determination of the boundary between the Oligocene and Miocene deposits in other areas of the Crimea–Caucasus region should be implicitly based on their correlation with the sections of the stratotype area.

Since the adoption of this resolution, certain essential features were added to the characteristics of the transition interval between the Oligocene and Miocene, which are as follows.

The Caucasian regional stage was determined as the most ancient part of the Miocene in the western Paratethys (Nosovskii and Bogdanovich, 1979) approximately conforming to the Aquitanian Stage of the West European Neogene scale; this circumstance was recorded in the summary volumes of *Stratigrafiya SSSR*, (1989, vol. I, p. 420).

The reference section of the Caucasian regional stage was comprehensively studied by using data derived from the parastratotype Well 4 drilled 45 km north of Tikhoretsk on the Stavropol uplift (Nosovskii, 1992).

The Oligocene–Miocene deposits of the Maikop Group were also studied on the basis of reference wells specially drilled on the Chongarskii Peninsula in the junction zone between the Crimean Plain and the Sivash area (Nosovskii, 1993).

The faunal assemblages from the Sivash beds of the Dnieper–Donets depression were revised, and the probability of their reference to the lower Miocene within the Caucasian regional stage was indicated (*Opredelitel' miotsenovykh...*, 1993).

The Nugri Formation of Ergeni, the mollusk fauna of which was used to establish the late Oligocene age of the Askaniya Formation in the northern Black Sea area (Veselov and Nosovskii, 1962), is now referred to the lower Miocene in the general correlation scheme as coeval with the Aquitanian (Popov *et al.*, 1993).

It is important to note that in the Mediterranean province, the search for the Oligocene–Miocene boundary involved a detailed study of the Lemme section in the Tertiary Peidmont basin, North Italy, which was subsequently recommended as the global boundary stratotype for these two Cenozoic units (Steininger, 1992).

Taking into account the data given above, it seems important to discuss the present state of the problem concerning the Oligocene–Miocene boundary in the Maikop Group by using the sections of the stratotype region in southern Ukraine as the most complete and well-characterized by organic remains at all levels of the section.

Two major circumstances should be considered in order to correctly place this boundary. First, the early Miocene age of the Aquitanian Stage underlying the Burdigalian is univrsally adopted. Second, the positioning of this boundary in the Crimea–Caucasus area should be based on the most reliable datum level, such as the first desalination stage of the Paratethys in the Oligocene, which was traced in Eurasia from the Alpine foothills in the west to the offspurs of the Pamirs and Tien Shan in Central Asia. The deposits of this so-called Solenovsk basin of southern Ukraine begins with the basal marl bed bearing ostracods (the Molochansk Formation). These deposits overlie the noncarbonate clays of the lower Maikop Group, thus indicating an alteration event in the abiotic medium, which changed the Rupelian type of fauna of the sea basin with normal salinity to the brackish-water assemblages at the beginning of the Molochansk and subsequent Serogoz time.

The Molochansk deposits contain occasional molluscan species *Ergenica cimlanica*, small cardioids of the genus *Cerastoderma,* and abundant ostracods *Disopontocypris oligocenica, Echinocythereis variatuberculata*, and *Loxoconcha nystiana.*

The intensive carbonate sedimentation of the Molochansk time was succeeded by accumulation of silty–sandy noncarbonate rocks of the Serogoz Formation. This event is recorded in a profusion of representatives of the Corbulidae family, among which two molluscan biozones are distinguished. The lower *Janschinella garetzkii* Zone includes *J. garetzkii, J. vinogrodskii* and *J. melitopolitana*, and also marks the first appearance of *Cerastoderma serogosicum.* The overlying *Corbula sokolovi* Zone begins at the first occurrence level of *Corbula* species attaining mass development in the second half of the Serogoz time along with similarly abundant cardioids represented by endemic species *Cerastoderma serogosicum* and *C. chersonense*. In addition, *Glycymeris pilosa lunulata, Polymesoda convexa, Siliqua taurica, Melanopsis hantkeni*, and *M. callosa* were noted at this level.

The same composition of faunal assemblages in a similar stratigraphic position was observed in the entire eastern Paratethys, from the lower reaches of the Dnieper River in southern Ukraine to the Transcaspian region. These are assemblages of the Tsraudon Beds in northern Osetiya, the Ikiburul' Beds in the Volga–Don interfluve, the *Corbula* Beds of eastern Georgia, the Tskhruta–Tsakhana Beds of the Akhaltsikhe depression, and the Tori and *Cerastoderma serogosicum* beds of Kyzylkum. The rocks of the Serogoz age in the Crimea–Caucasus region contain biocoenoses with mollusks of the Serogoz type. The coenoses laterally substitute each other and occasionally contain accessory elements of typical marine fauna. The appropriate examples are the Serogoz, Kostromka, Tsraundon, South Ergenei, Akhaltsikhe, Aralian, and other biocoenoses.

Analogous sediments containing fauna of the Molochansk–Serogoz type and occupying a similar stratigraphic position are known in the Central Paratethys (Baldi, 1989): the Transilvania basin (Buzash Sand and Ileand Clay), Hungary (*Sirenia* Beds and clay of Tard), Bulgaria (Ezerivo Formation), Austria (the Washberg zone), and Slovenia (Sottska Beds).

Apparently, their analog in the Cis–Alpine basin is the *Sirenia* Marl of the Mainz basin, where it is referred to the Chattian Stage on the basis of brackish-water mollusks and ostracod assemblages. This unit underlies lacustrine sediments of the Aquitanian Stage with *Helix ramondi* (Zöbelein, 1960).

The indicated endemic fauna of the brackish-water mollusks from the eastern and western Paratethys is unsuitable for an unambiguous age determination of host deposits. It is, therefore, necessary to take into account not only the ecological features of the fauna, but also its position in the section, the presence of other groups of organic remains, and also geodynamics and paleogeography of the basins.

The composition of ostracods from the Serogoz beds of the Black Sea region is studied in detail by Sheremeta (1969), who has shown that their species (about 14 taxa) are as diverse as in the Chattian of the Mainz basin and Switzerland, and also as in the upper Oligocene assemblages from the northern German lowland. The similarity of ostracods allowed Sheremeta to refer the Serogoz deposits of southern Ukraine to the upper Oligocene (the Chattian Stage).

I am inclined to associate the formation of this brackish-water semiclosed basin with the very significant Chattian regression in many regions of the world, when a sudden drop of sea level was about 100 m and more, as indicated by some researchers (Vail and Mitchum, 1979; Ronov, 1994). There is probably no other explanation to this important geohistorical event associated with the eustatic oscillation of sea level that caused sudden changes in the biotic and abiotic environment of Eurasia at the end of the Oligocene.

The period of time subsequent to the late Oligocene accumulations of the brackish-water Serogoz sequence corresponded in the south of Ukraine to a considerable transgression of a sea basin with normal salinity that afterward left the so-called Askaniya deposits.

The Askaniya transgression is recorded not only in the south of Ukraine, but in other parts of the eastern Paratethys as well. In all the localities, it left deposits with a peculiar assemblage of polyhaline mollusks, foraminifers and ostracods of the transitional Oligocene–Miocene type. This assemblage is intermediate between faunas of the Boreal and Mediterranean bioprovinces and may be interpreted as the late Oligocene or early Miocene in age, depending on which faunal assemblages it is compared with in the Crimea–Caucasus region, or in the West European sections. The problem is complicated by the fact that this part of the section, known as the middle Maikop subhorison in the Crimea–Caucasus area, is still inadequately correlated within the eastern Paratethys, though a rather sound faunal correlation is suggested by Korobkov (1974), and the Kalmykian regional stage is identified in this part of the section by Popov *et al.* (1993).

The Askaniya Formation was initially referred by Veselov and Nosovskii (1962) to the upper Oligocene on the basis of mollusks identical to these fossils from the Nugri Formation of Ergenei described by Liverovskii (1939). Later, the age of the Askaniya Formation was determined as early Miocene on the basis of the paleohydrological method of stratigraphy suggested by Zhizhchenko (1964) and on the composition of ostracods from these deposits (Sheremeta, 1969). It is remarkable that the age of the Nugri Formation of the Volga–Don interfluve is regarded in one of the latest publications as a part of the Aquitanian interval of the early Miocene (Popov *et al.*, 1993).

At this point, we meet again the problem of the Caucasus regional stage established in the stratotype section of the Kuban River near Cherkessk (Nosovsky and Bogdanovich, 1979). This stage includes the Alkun and Zelenchuk formations with septarian beds at the base, together with the lower half of the Karadzhalga Formation.

According to Bogdanovich, the Caucasus regional stage bears two foramineral assemblages: the lower one with typical marine species *Bolivina goudkoffi caucasica* and *Virgulinella neobuliminiformis,* and the upper one with *Uvigerinella californica,* which is considerably impoverished and lacking several forms of polyhaline benthos.

These foraminiferal assemblages are genetically uniform and span the gap between the Ol'ga Formation and Solenovsk Horizon (the Chattian), thus dividing the Caucasus regional stage into two substages.

The lower *Bolivina goudkoffi* assemblage is traced as yet only in the Crimea region of southern Ukraine. It is detected at the base of the upper Kerleut beds of the Crimea sections in the Indol depression southeast of the Vladislavovka railway station and north of Feodesiya. According to the determinations of I.V. Venglinskii, the remains of small foraminifera at these localities belong to the benthic species *Bolivina goukoffi* and to the planktonic assemblage of *Globigerina ciperoensis, Gl.* cf. *officinalis, Gl.* ex gr. *danvilensis,* and *Gl. juvenilis.* This part of the section also demonstrates the coexistence of *Spiroplectammina caucasica* and *Sp. terekensis.*

In the Black Sea region, the lowermost part of the upper Kerleut Beds corresponds to the Askaniya Formation, well known by its foraminiferal assemblage of the *Spiroplectammina terekensis–Sphaeroidina variabilis* Zone. The concurrence of basal upper Kerleut Beds of the Crimean Plain and Askaniya Formation of the Black Sea region is confirmed not only by foraminiferal assemblages, but likewise by the consecutive tracing of borehole sections along the profile extending from the Crimean Plain region north of Dzhankoi toward Askaniya Nova in the northern Black Sea area. This correlation is also controlled by the species composition of peculiar malacofauna from this part of the section in both regions, most frequently represented by *Nuculana (Saccela) gracilis, Palliolum (Delectopecten) simile, Plagiocardium abundans,* and some other forms. Zosimovich (1981) considered the Askaniya and Sivash molluscan assemblages, the latter from the Dnieper–Donets depression (DDD), as coeval and corresponding in age to the late Oligocene. The latest revision shows, however, that the Sivash fauna of DDD corresponds to the Caucasian regional stage of the lower Miocene (*Opredelitel' miotsenovykh...,* 1993). A similar conclusion about the Miocene age of the Sivash fauna was also given in another publication (Nosovskii, 1995).

Upward the section, the Askaniya clay deposits are very gradually replaced by the silty–clayey rocks of the Gornostaevka Formation, which is most fully represented in the lowest part of the northern Black Sea depression at the latitude of the Karkinitskii Bay.

The deposits of the Gornostaevka Formation are studied in detail by two wells specially drilled for stratigraphic purposes in the Chongarskii Peninsula in the northern part of the Black Sea region (Nosovskii, 1993).

As shown by borehole data, the deposits in the lower part of the Gornostaevka Formation are 118 meters thick and contain siliceous organisms: radiolarians, sponge spicules, and diatom remains. This unit is barren of other fossils, but the overlying middle part of the Gornostaevka Formation, 140 meters thick, is rich in faunal remains. According to characteristic benthic foraminifers, this fauna is classified as the "nonionid" assemblage. In aforementioned and other wells nearby, the assemblage includes *Porosononion dendriticus, Nonion granosus, Melonis dosularensis, Elphidium onerosum, Heterolepa ornata,* and remains of mollusks *Venus multilamella, Corbula gibba, Parvicardium abundans,* and *Cerastoderma prigorovskii.* The upper part of the Gornostaevka Formation is 45 meters thick and contains foraminifers *Uvigerinella californica* and *Spiroplectammina caucasica*; glauconite nuclei of these rocks contain shells of *Bolivina* and *Pseudoparrella.*

This succession of the Gornostaevka faunal assemblages is characteristic of all borehole sections in the area of the northern Black Sea region and Crimean Plain.

It should be noted that the horizon with siliceous organisms is traced not only in the deep wells drilled in the northern part of the Black Sea region, but also in boreholes of the Crimean Plain and Indol depression of the Kerch Peninsula. In the Crimea–Caucasus region, this horizon probably represents an interregional unit reflecting peculiar ecological factors, because the spongolite deposits are also observed in the western Ciscaucasia, in the Voskovogorsk and Assinsk formations of the middle Maikop Subhorison of Kuban, and in northern Osetiya, (Bogdanovich, 1959; 1960).

In these regions of southern Ukraine, the Gornostaevka Formation is the last in the succession of the Caucasian regional stage; it is overlain by terrigeneous rocks beating *Neobulimina elongata* (the Chernobaevka Formation) and is referred to the Sakaraulian Stage of Eastern Paratethys.

Palynomorphs from the Serogoz, Askaniya and Gornostaevka formations of the Chongarskii section were studied by Shchekina (1969; 1970), who found them to be similar in many features to the middle Maikop pollen assemblage of Crimea and northern Caucasus and to the Chattian–Aquitanian assemblages of the Central Paratethys. Moreover, spores and pollen from the Chernobaevka Formation resemble their assemblages from the upper Maikop Subhorison of Kuban, Stavropol and, consequently, from the Burdigalian of Western Europe.

Among other groups of organic remains from the Askaniya and Gornostaevka deposits of southern Ukraine, the ostracods and peridinians should be noted. In contrast to the Molochansk–Serogoz beds, ostracods are less frequent in the overlying strata. They were recovered from basal beds of the Askaniya Formation. This assemblage of 25 species, some of which are inherited from the Serogoz deposits, is rich in forms that became flourished in the Miocene, and Sheremeta (1969) has concluded that the assemblage is the early Miocene in age. Slovakian paleontologist P. Jiříček (1975) arrived at the same conclusion after his analysis of the ostracod fauna in this region. He emphasized that in the south of Ukraine, this fauna can be considered as corresponding to *Hemicyprideis helvetica* and *Neomonoceratina helvetica* superzones, and the analyzed deposits can be analogs of the Aquitanian Stage.

Peridinians from these beds are as yet insufficiently studied and inadequately described (Andreeva-Grigorovich, in *Stratigrafiya SSSR*, 1986; vol. 2, p. 360). However, the extensive distribition of the representatives of the *Deflandrea* genus in the Askaniya and Gornostaevka beds indicates the probable concurrence of these deposits, the more so as their appearance and flourishing was also observed in the analogs of the Aquitanian Stage, namely in the global Lemme stratotype of the lower Miocene in the Tertiary Piedmont basin (Steininger, 1994). This circumstance is favorable for further detailed analysis of distribution patterns of this group of organisms in the Crimea–Caucasus region.

On the basis of the described materials, the deposits of the Askaniya and Gornostaevka formations in the Black Sea region should be regarded therefore as exemplifying a single and continuous lower Miocene sedimentation cycle corresponding to the Caucasian Stage of the Eastern Paratethys.

As for the paleogeographic analysis of the sea basin of the Caucasian time, it is necessary to rule out the junction of Tethys with the Polar basin through the Turgai trough, the closing of which occurred in the middle of the Priabonian (Tavdian) about 37–40 Ma ago. The migration routes of respective fauna from the northern bioprovince likewise cannot be traced in the area of the East European platform, including Poland.

According to my views, the considered fauna of the Eastern Paratethys experienced a decisive influence of Western Paratethyan communities, whose representatives migrated through the Carpathian region and a strait located in the Lower Rhine graben in Western Europe; this strait apparently joined the Boreal basin of the Norwegian–Greenland Sea with the Paratethys. This is confirmed, among other factors, by the composition of the Toeroekbolinta fauna in northern Hungary and Kovachov fauna in southern Slovakia from basal *Glycymeris* Sands and overlying *Turritella venus* Beds. These faunas contain a considerable amount of Miocene faunal elements of the Atlantic and Mediterranean bioprovinces and also the relict Boreal species (Senesh and Tsikha, 1975). Therefore, if the Euxinian–Caspian area was the radiation center of the Serogoz brackish-water fauna in the Paratethys, then the polyhaline Caucasian (Askaniya–Gornostaevka) fauna could migrate from the Boreal province of the Northern Sea through the strait located in the Lower Rhine graben. Taking into account this paleogeographic situation, I consider it necessary to keep it in mind when revising age of a number of localities with polyhaline molluscan fauna in the Eastern Paratethys, since previously this fauna was usually correlated with that of the Chattian Stage in Northern Europe.

To sum up the materials discussed above, I should point out the following items.

(1) In order to draw the boundary between the Oligocene and Miocene in southern Ukraine, it is necessary to proceed from the established fact of the vast Molochansk–Serogoz regression that occurred at the end of the Oligocene (Chattian time).

(2) The Caucasian Stage in the Eastern Paratethys begins with a large transgression, which marks the lower Miocene boundary at the base of the Askanian deposits in southern Ukraine.

(3) The analysis of paleogeographic environment at the Oligocene–Miocene boundary and the composition of respective faunal assemblages provide means for tracing the migration paths of the brackish-water fauna from a semiclosed late Oligocene basin of the Eastern Paratethys into its western part and, at the commencement of the early Miocene, from the Boreal area of the Northern Sea to the Paratethys.

(4) The available paleontological data for the age interval under discussion imply that the suggested correlation will be substantially improved after a more detailed investigation and description of absolutely all recovered groups of the organic remains.

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Precambrian Stratigraphy as Discussed at the Conference on Correlation of Fennoscandian Geological Complexes

V. Z. Negrutsa

Geological Institute, Kola Research Center, Russian Academy of Sciences, Ul. Fersmana 14, Apatity, 184200 Russia

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The first international conference "Correlation of Fennoscandian Geological Complexes" was held in St. Petersburg September 8–11, 1996, and was hosted by the State Mining Institute.[1] The Conference was organized by the following institutions: the Committee of the Russian Federation on Geology and Exploring of Natural Resources, the Northwestern Regional Geological Center, the Mineral State Company, the St. Petersburg State Mining Institute. It was supported by the Russian Foundation for Basic Research (project no. 96-05-74049), "Karelgeolcom," "Murmangeolcom," the Institute of Precambrian Geology and Geochronology of the Russian Academy of Sciences, the Geological Institute of the Kola Research Center, the Institute of Geology of Karelian Research Center, the Interdepartmental Council for Precambrian Petrology and Mineral Resources, and the Petrographic Committee for the Northwest of the Russian Federation.

The program of the conference included 60 oral presentations and 46 poster reports. It consisted of five sections: (1) geology of the Ladoga–Bothnic Zone (6 contributions), (2) correlation of Fennoscandian ore-magmatic systems, (10 contributions), (3) isotopic-geochronological investigation of the Fennoscandinavian Shield (7 contributions), (4) regional and interregional correlation schemes (14 contributions), and (5) deep structure, geodynamics, and magmatism of Fennoscandia (14 contributions).

The conference focused on problems of the correlation of Archean and Proterozoic magmatic and metamorphic ore-bearing complexes, which were long studied by joint Russian–Finnish and Russian–Norwegian groups of specialists. Stratigraphic aspects of these wide-range problems were analyzed by a rather small team (about 20 persons) of specialists, mostly Russians. Here in Section 4, 9 out of 14 included in the program contributions were presented and discussed. V.A. Prozarovskii (St. Petersburg State Univ.), A.M. Akhmedov (All-Russia Institute of Geology, VSEGEI), V.Z. Negrutsa and T.F. Negrutsa (Geological Institute of the Kola Research Center, RAS) dwelt upon methodological issues of Precambrian stratigraphy. Presentations by Yu.B. Bogdanov (VSEGEI), M.A. Korsakova, and N.M. Ivanova (State Geological Enterprise "Sevzapgeologiya"), and that of V.Z. Negrutsa were devoted to the elaboration of uniform stratigraphic schemes and legends for maps of different scales. V.A. Melezhik (Geological Survey of Norway, Trondheim) and A.E. Fallik (Scottish Universities Research and Reactor Center, SURRC, Glasgow), coauthored by P.V. Medvedev and V.V. Makarikhin (Institute of Geology, Karelian Research Center RAS), and R.E. Prilutskii (VSEGEI) reported on carbon and oxygen geochemistry and isotopic characteristics of carbonate sections; A.V. Savitskii outlined the systematics of isochron dating of uranium minerals.

V.A. Prozarovskii noted a fundamental significance of stratigraphy for all geological activities, reviewed its major objectives (such as the subdivision and correlation of sections), and considered three categories of stratigraphic scales (local, regional, and general). He substantiated the idea that the general scale should be based on one of regional scales (or composed of several scales) accepted as a standard of the ideal crustal section without repetitions and overlaps. He also concluded that Global (Regional) stratotype section and points should be chosen for global (regional) boundaries of each subdivision of the regional and general chronostratigraphic scales. According to this communication, the historical–geological approach to the problem inevitably results in obtaining the diachronous boundaries, whereas isotopic dates appear to be rather controversial. Therefore, fixed reference points ("golden spikes") should be chosen in sections with conformable bedding.

A.M. Akhmedov suggested that regional and interregional correlations of the Early Proterozoic sections should be done with due regard to: (1) tectonic typization of structures, (2) analysis of formations sensible to and indicative of the state of oxigen potential in the Earth's upper crust, (3) epochs of maximum bioproductivity. The author noted that certain interrelated criteria taken together provide grounds to identify the following periods of the Lower Proterozoic: (1) the global Sariolian (2500–2300 Ma ago) and local Kalevian epochs of glacial lithogenesis; (2) the global Late Jatu-

[1] *Korrelyatsiya geologicheskikh kompleksov Fennoskandii. I Mezhdunarodnaya konferentsiya, 8–11 sentyabrya 1996* (Correlation of Geological Complexes of Fennoscandia. The First International Conference, September 8–11, 1996), Abstracts of Papers (Russian and English), St. Petersburg: Mineral, 1996.

lian (2200–2100 Ma ago) and local Vepsian epochs of oxidized evaporite sedimentation; and (3) the global Ludicovian (2100–2000 Ma years ago) and local Kalevian epochs of development of anoxic basins. Moreover, the Late Jatulian and Ludicovian are epochs of widespread aerobic and anaerobic biocenosis, respectively. The boundary between them (its age is 2100 Ma) is the main correlation level of the Lower Proterozoic.

Noting the multivariant and schematic character of geochronometric scales and using age dates as markers of definite geological events, irrespective of the method and object of isotopic chronometry, V.Z. Negrutsa and T.F. Negrutsa supported the idea of creating a general cyclorhythmochronometric scale for Fennoscandia, which depicts up to 9 hierarchical subdivisions (spanning intervals from more than 1.0 Ga to less than 20 Ma). They stressed the fundamental importance of settling two interdependent problems: (a) the elaboration of a single network of stratotype sections and boundary standards for regional and interregional strata of Fennoscandia, and (b) the comprehensive investigation of geological events and isotopic–geochemical features of stratotypes selected in conformable sections. The speakers believe that there is a sufficient and well-substantiated basis for solving these problems.

Yu.B. Bogdanov pointed to a fundamental significance of the stratigraphic scheme of the Karelia–Kola region, adopted by the Regional Interdepartamental Stratigraphic Committee for northwestern Russia and approved by the All-Russia Interdepartmental Stratigraphic Committee in 1989 for elaborating the legends of detailed geological maps of the Kola–Karelia series and to maps at the scale 1 : 500 000 for the entire shield area. Bogdanov suggested that the principle criterion of regional correlations is the isotopic age of rocks obtained by U–Pb methods on zircons. Bogdanov, in cooperation with O.A. Levchenkov *et al.*, dated, for the first time, the age of zircon from quartz porphyries of the Hogland Island (the Gulf of Finland). The obtained value 1638 ± 3.8 Ma confirms the Early Riphean age of the Hoglandian. The researchers also showed that the Sumian Ozhiyarvi Formation (its zircon age is 2420–2440 Ma) is separated from the underlying volcanic rocks of the Tungud Formation by conglomerates, the base of which, most likely, coincides with the Archean–Proterozoic boundary.

M.A. Korsakova and N.M. Ivanov reported on their experience in correlating plutonic, metamorphic, and metamorphosed sedimentary–volcanic rocks in the Karelian region as significant for the legend of a new generation of detailed geological maps of Karelia. According to the program, the same problem should have been discussed by A.M. Remizova in her report on the Kola Peninsula, but, regretfully, this communication was omitted. The abstract of this paper presents results concerning the successive elaboration of local stratigraphic scales, their intercorrelation, and correlation of local subdivisions with those of the regional and general stratigraphic schemes. Attention was also focused on some unsolved problems, particularly, on the space and time relations of variously metamorphozed endogenic and exogenic rocks and dated intrusions with stratigraphic units of regional, general, and, in many cases, local scales.

V.Z. Negrutsa dealt with the evolution of views on geochronology, and with elaboration of the modern stratigraphic scale of the Precambrian of the Eastern Baltic Shield, which is the type scale for northern Eurasia. He emphasized that stratigraphic schemes of the Lower Precambrian in the eastern Baltic Shield, worked out in the 1970s and approved by the Interdepartmental Conference on Stratigraphy in 1982, are based on standard established sections (stratotypes), for the first time, for the regional and interregional stratigraphic subdivisions. Using the chronostratigraphic correlation of Lower Proterozoic sections, Negrutsa demonstrated the lateral variability of facies composition of identified subdivisions and the dependence of boundary ages of lithostratigraphic subdivisions on the completeness of geological of sections, and also discussed as inevitable the multivariant character of geochronological periodization of the Precambrian. The author suggests that this is a result of a situation when stratigraphic gaps between lithostratigraphic subdivisions change in a lateral direction and the unit boundaries are unfixed in continuous sections.

V.A. Melezhik and his colleagues (A.E. Fallik, P.V. Medvedev, and V.V. Makarikhin) outlined results of their research on carbon and oxigen isotopes in stromatolitic dolomites of the Tulumozero Horizon (Paleoproterozoic) of the Onega depression (Karelia). The authors suggested that the positive $\delta^{13}C$ anomaly (+5...+17‰) recorded in this horizon is of a worldwide scale. Explaining it, they discussed a tectonic model, a model of the layered ocean, and a model of an evaporite stromatolite basin.

P.E. Prilutskii expressed the opinion that the theoretical base of the Precambrian carbon-isotope hemostratigraphy is controversial. He proposed an original method of evaluating the primary carbon-isotope composition in sedimentary carbonates, taking into account the distortion of the carbon-isotope system in the course of postsedimentary transformations.

A.V. Savitskii *et al.*, presented previously unpublished results of U–Pb dating for more than 40 uranium-ore formations of Karelia and the Kola Peninsula. The speaker concluded that, in the context of these data, the chronological limits of superhorizons of the regional stratigraphic scale are as follows: 2500–2400 Ma for the Sumian; 2400–2350 Ma for the Sariolian; 2350–2200 Ma for the Jatulian; 2200–1970 Ma for the Ludikovian; 1970–1930 Ma for the Kalevian; and 1930–1860 Ma for the Vepsian. The interval of 1860–1650 Ma corresponds to a pulsation phase of the regional tectonic activity with the intensity maximum in the period of 1780–1730 Ma. Other recorded impulses of

tectono-magmatic activity took place 1200, 1000, 800–600, and 400–200 Ma ago.

The posters presented by T.F. Negrutsa, A.V. Sochava, and A. Sidletska and published abstracts (A.S. Voinov, V.Ya. Gor'kovets, and M.B. Raevskaya, T.E. Magnitskaya *et al.*, A.P. Svetov and A.V. Pervukhina, A.I. Svetova and S.A. Svetov, A. Sil'vennoinen *et al.*, Chikirev, I.V., I.V. Chikirev and A.A. Basalaev, A.S. Shchelemotov, and V.P. Chuikina) were devoted to research of certain section types and reference localities of Archean and Proterozoic deposits. Certain problems significant for stratigraphy were also discussed in papers concerned cartography, isotopic–geochemical and isotopic–geochronological investigations of metamorphic complexes, and with correlation of magmatic, metasomatic, and other events, including the epochs of metallogenesis.

The comparative analysis of all these contributions, which was partially debated at the plenary session (speakers A.M. Akhmedov, Yu.B. Bogdanov, A.B. Bulavin, S. Dali, D.A. Dodin, G. Kautskii, K. Korsman, E.I. Lukonen, E.B. Nalivkina, V.Z. Negrutsa, T.F. Negrutsa, V.V. Proskuryakov, S.I. Turchenko, N.B. Filippov, V.I. Shul'diner, and E. Ekdala) and summarized in the memorandum adopted by the conference, emphasizes the following points.

1. Over the recent years, Fennoscandian geologists have realized that the more realistic and detailed geological works (cartographic in particular), the more reliable and detailed their background. The necessity of using methods and experience gained in Phanerozoic stratigraphy is increasingly felt by Fennoscandian geologists by solving problems of the Precambrian periodization. Essentially important is the goal to elaborate a network of stratotype sections and boundary stratotype points of local, regional, and interregional stratigraphic subdivisions, and to promote complex research aimed at selecting international standards and their complex analysis. The modern stratigraphic base of geological research is inadequate, since criteria for the subdivision and correlation of local sections are poorly elaborated, the stratigraphic position of isotopically dated magmatic and metamorphic complexes is imperfectly known, and conceptual approaches and nomenclature are diverse.

(2) The Precambrian chronostratigraphic scale of Karelia and the Kola Peninsula, which is based on stratotypes and the facies–paleogeographic subdivision and correlation of real sections, may serve as a reference framework for the entire Fennoscandia. When coordinated with this scale, stratotypes of general subdivisions and their multidisciplinary study will bring us to a new level of geological metrology and investigation of space-and-time and cause-and-effect relations between the processes of development of external and internal geospheres.

(3) The main objective of the further elaboration and detalization of the Precambrian stratigraphic scale for Fennoscandia is to provide a better methodological basis for the regional Precambrian stratigraphy and to elaborate a unified system of stratotype sections and standard boundaries of stratigraphic subdivisions for further detailed investigations.

The brief memorandum of the conference informs that 105 experts from Russia, Finland, Sweden, Norway, Ireland, Scotland, and USA participated in the session. It is also noted that the problems of geological correlation considered at the conference are principal subjects for further joint activities on the preparation of cartographic materials, prospecting and exploration of raw mineral materials. Further international cooperation and efforts of scientific and industrial organizations are needed to settle the problems we are facing. Participants of the conference considered it desirable to hold regular meetings on correlation, geophysics and geochemistry of Precambrian geological complexes of Fennoscandia. The next conference is scheduled for 1998.

Reviewers Yu.R. Bekker and M.A. Semikhatov

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